RELATIONS BETWEEN WATER AND SOIL

by

T. J. MARSHALL

Technical Communication No. 50
Commonwealth Bureau of Soils
Harpenden

COMMONWEALTH AGRICULTURAL BUREAUX

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(Division of Soils, Commonwealth Scientific and Industrial Research Organization, Adelaide, Australia)

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PREFACE

This account is written from the standpoint of the agricultural use of soils. It touches also at some points on work done in other fields concerned with fluids in porous media. Soil is hardly to be considered as a single type of porous material since it can include inert sands on the one hand and reactive clays on the other. Hence some of the work done by the petroleum technologist on fluids in porous rocks, by clay technologists on water in clays, and by engineers concerned with soil as a foundation material is relevant to the present discussion.

Many terms have by now come into use in the general field of research on fluids in porous media. In facing the problem of what terms to use, I have been helped by reports on terminology published in the Proceedings of the Soil Science Society of America in Vol. 16 (1952) pp. 85-88, Vol. 20 (1956) pp. 430-440, and Vol. 22 (1958) p. 270 and by discussions on terminology of soil moisture and salinity initiated (but not yet published) by the International Committee for Horticultural Congresses.

I should like to thank Mr. G. V. Jacks of the Commonwealth Bureau of Soils and Dr. H. L. Penman of Rothamsted Experimental Station, England, for early discussions on the scope of this account and I am indebted to Dr. Penman and Dr. J. N. Luthin (while on a visit from the University of California), and Mr. J. W. Holmes, Mr. C. G. Gurr, Dr. E. L. Greacen and Mr. G. B. Stirk (Division of Soils), for reading some parts of the manuscript. I owe much to colleagues of the Division of Soils for many discussions and to Dr. J. P. Quirk (University of Adelaide) and Dr. K. L. Sutherland (Division of Industrial Chemistry, C.S.I.R.O.) for discussion of some of the material on sorption. Translations of a number of the papers referred to here were made by the Translation Section, C.S.I.R.O., Melbourne, and by the Commonwealth Bureau of Soils.

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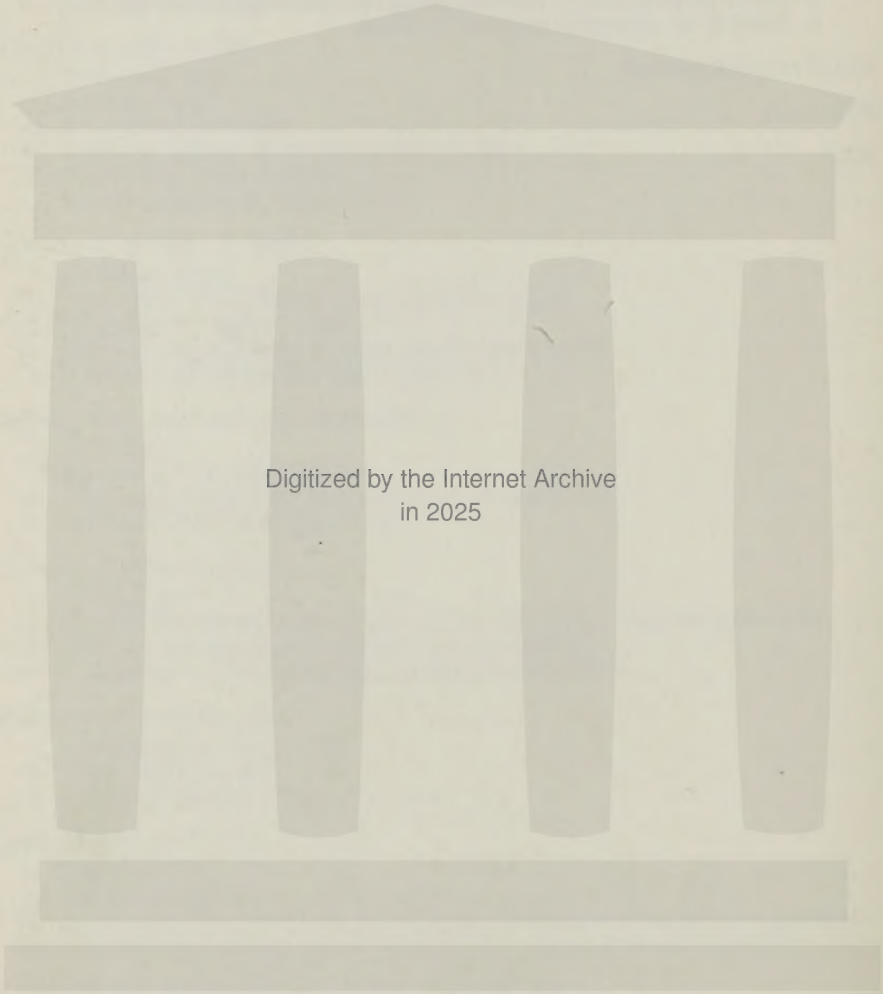

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1. SOIL AS A STORE FOR WATER

1. Introduction: Water and Some of the Properties of Soils

Over a great part of the land surface of the earth, water limits plant growth either because there is too much or too little of it in the soil. The extreme cases are in swampy and arid areas where if crops are to be grown at all they have to depend upon drainage or irrigation. Between these extremes lie the established agricultural lands many of which can be improved by having less water or more in the soil at certain times of the year. A great deal is done to control water in agriculture and, if both water and soil are to be put to the best use, a full understanding of soil-water relations is necessary. Much of the research on retention and movement of water in soil and the use of water by plants is done with this objective.

Water has notable effects upon the soil itself and these may indirectly affect plant growth also. It plays a key part in soil formation. The processes of weathering and profile development (involving organic as well as inorganic parts of the soil) are greatly affected by water. The movement of materials dissolved in water results in horizons of loss and of accumulation which give the soil a distinctive profile and may brand it as fertile or infertile for plant growth. Water is also a responsible agent in soil erosion. Many of the physical properties of soil, such as its structure, strength, and swelling, are greatly affected by it so that soil-water relations play an important role in tillage and in the design and performance of roads and buildings.

The behaviour of water in soil is affected by the size of the particles and the way in which they are arranged. Generally the greater the clay content, the more water will soil retain at a given stage of draining or drying, but the type of clay mineral and the nature of the exchangeable cations also affect retention and movement. The particles may be openly or closely packed together and so leave more or less pore space between them for water and air. They form the porous matrix in which the water is embedded and the way in which they are arranged so as to leave much pore space or little, large pores or small, is clearly just as important in water relations as the size and kind of the particles. The total pore space (the volume fraction occupied by water and air) is fairly easy to determine but the pore sizes have been more troublesome. Water will move through, or drain out of, large pores more easily than small, and so size of pores has to be considered for water storage and movement as well as amount of pore space. The classical approach to pore size was through its particles. Workers trying to find some basis for relating soil properties to permeability, at first determined the geometry of the pore space in ideal systems of packed spherical particles. Later they obtained an effective tube size from the ratio of porosity to surface area of the particles. By this means, irregularly shaped particles could be dealt with; but the method failed when there was a larger range of size and it is unsuited to soils. With the development of methods for finding how water content changes when a suction is applied to the water in the soil, a means ultimately became available for measuring the size distribution of pores in non-shrinking soils.

The theoretical background for water retention, movement and storage in unsaturated soil was originated more than 50 years ago by Buckingham but advances in these, as in most lines of soil-water investigation, were greatly accelerated by developments in techniques for measuring suction which came much later. These have led to a better understanding of the processes by which water becomes distributed through soil and is made available to the plant.

II. Gains of Water to the Store

Soils gain water from rain and irrigation and lose it by drainage, by transpiration from plants, and by evaporation directly from the soil. Some of the processes by which water is gained or lost can be controlled. Control over water entry is mainly through irrigation but man also undertakes treatments which are designed to increase or reduce the amount of rain water entering directly into soil. Soil and water conserving practices which improve soil permeability or which hold the water ponded on the surface until it soaks in, are examples of what is done to increase intake and so reduce loss of water which would otherwise run off the surface.

The amount of water retained in soil which has drained for about two days (while covered to prevent evaporation) is known as the field capacity. This is not a precise quantity but it serves the useful purpose of providing an approximate upper limit to the amount stored for subsequent use by plants. The lower limit is the permanent wilting point and the difference is the range of available water for plant growth. This definition of capacity for water storage has been widely used. In irrigation practice in particular, it has provided a useful basis for determining when water is needed and the amount required to recharge the store. Excessive irrigation, much beyond that amount, is wasteful of water and damaging to the soil.

In order to know at any time the actual stage of drying reached within these limits and so decide how much and when water is needed, it is usual to sum up the position qualitatively from local experience with soils, plants, and climate. Quantitative methods are also in use based on plant behaviour (e.g., changes in fruit size), evaporation measurements, or soil measurements. Soil measurements are of two kinds. If it is desired to know the actual amount of water to be added to the surface in centimetres or inches, in order to recharge the store for a particular soil, then a measure of the water content (as the amount of water in a given volume of soil) may be needed. However, the best common guide to the stage of drying is given by the suction of the soil. Suction is a term referring to the force with which water is held in the soil. For example, if water in a porous cup is brought in contact with moist soil, then the suction which is needed to keep it from entering the soil is equal to the suction of the soil. Both the terms suction and water content will be dealt with more fully in the next section.

In large irrigation areas when watering times are controlled on a regular roster, the watering schedule is not very flexible; but the smaller areas and especially the one-farm units with independent water supply have more latitude. In all cases there is the desire for answers to the question of water needs and a keen demand exists for methods for measuring the changing suction or water content for practical agricultural and field experimental work. A number of methods have been introduced to obviate sampling, and instruments such as tensiometers and gypsum blocks have proved useful for measuring suction in the field. Others with promise, such as the neutron moisture meter, are becoming available for measuring water content. Also, provided plants suffer no restriction in the supply of water to their roots, the amount lost by transpiration can be calculated from meteorological data. If some known starting point such as field capacity is taken, then the deficit of water at any later time can be calculated by this means if other losses can be disregarded. Advantages and limitations of various methods for determining suction and water content in the field will be discussed in a separate section.

III. Losses to the Atmosphere

Evaporative losses can be most easily controlled by removing all vegetative cover. The most widespread use of this means of control is in the fallow periods common especially in cereal farming. There are varying reasons for fallowing but in some regions the storage of water for the next crop is the main one. Much work has been done on treatments to conserve water. Paper, foil, or plastic sheets and various other materials can be used to cover the ground and reduce water loss. A comprehensive review of these and soil treatments has been made by Jacks, Brind and Smith (1955). Treatment of the soil itself usually depends first of all on the removal or prevention of weeds so that there will be no unnecessary withdrawal of water by plants. Loosening of the soil surface can reduce liquid flow to the surface and so create a dry barrier which reduces evaporation. Ordinarily the top soil will dry out any way and the protection will thus be provided in dry weather without loosening the soil. Water is lost from a bare soil by evaporation at approximately the same rate as from an open water surface so long as the

soil surface remains wet. When the surface dries out due to the rate of evaporation exceeding the rate of supply from below, evaporation decreases markedly. Penman (1941) considered that when a wet soil is subjected to conditions which induce a rapid rate of drying, the total water lost over a period may be less than would be the case if drying conditions were less severe. With rapid drying a protective dry mulch is quickly formed. He has suggested that, in view of the continued loss with slow drying, cultivations in spring might be effective in reducing loss of water from bare soil. They are ineffective in the summer except by way of weed control. In later work, Gardner (1958) and Gardner and Fireman (1958) show that the rate of loss depends on the rate of supply from below in the liquid phase unless the meteorological conditions for evaporation are themselves limiting.

Lemon (1956) has used surface active agents which, by reducing the surface tension of the water, cause less water to be retained in the surface soil after an irrigation. Hence the dry zone will be established sooner and with less loss of water than otherwise. Results were said to be promising but unproved on a field scale. Earlier Russian work using soap for this purpose, is quoted by Lemon. Russian workers have paid a lot of attention to surface treatments of soil structure to reduce movement in both liquid and vapour phases. Koliasev (1941) and Kolesnik (1948) found that a layered structure of alternate loose and compact soil was effective in reducing the loss of water from the top metre of soil. In Kolesnik's experiments, soil treated in this way conserved more water over the summer and at the time of seeding had 4 cm more to a depth of 1 metre than one which was loosened but had no compact layer. The purpose of the compact zones was to reduce vapour movement and that of the loose zones to reduce liquid movement towards the surface. Burov (1952) showed that the amount of water lost to a depth of 1 metre increased with size of the structural aggregates (and the amount of pore space made up of larger pores) because of wind entry when the pores are too large. Also if they are too fine, liquid loss (and cracking) can become serious. An optimum size of aggregate to prevent evaporation was found by Burov (1954) in field experiments to be 3-0.25 mm. Lumpy (>10 mm) and fine (<0.25) aggregates increased the losses under field conditions and these fractions were best kept below 5 to 10 per cent by weight for Trans-Volga chernozem. Burov suggests that this type of finely crumbed structure could be brought about by pasture leys and proper tillage.

A change in the annual amount of water used by vegetation can have widespread effects on water storage, run-off, and soakage to aquifers. In some circumstances, the burning off of perennial native shrubs so that annual grasses can take over, has been advocated. In California, Veihmeyer (1953) has shown that, in a climate with cool wet winters and hot dry summers, the grasses do not dry the soil out so thoroughly (except near the surface) and there is more water remaining in the soil when the following wet season begins. The practice may be opposed on other grounds such as erosion (*see* discussion of Veihmeyer's paper, *Trans. Amer. Geophys. Un.* 35, 652-658) but there is no doubt that the principle is sound so far as the water regime alone is concerned. The effects of clearing perennial native shrubs and trees and substituting annual winter-growing grasses, crops, and fallows upon local hydrology can sometimes be unwelcome. Teakle and Burvill (1938) have studied one case of importance in which clearing of native vegetation following settlement for agriculture has resulted in increased seepage of water to the valleys. In this case the soils, were naturally fairly saline and the seepage has resulted in serious salting of valleys in a large agricultural region in Western Australia. Similarly Wilde *et al* (1953) found that the clear cutting of aspen stands on podzolized soils in northern Wisconsin caused semi-swamp conditions to develop.

IV. Removal by Artificial Drainage

Excess water in soil is removed artificially by subsoil drainage to a depth chosen on the basis of the type of soil, crop, and climate and the economics of drain construction. In humid areas the depth is often of the order of 100 cm (or less) but in irrigated arid country it is commonly deeper and if soils are permeable to that depth, tile drains are commonly placed at a depth of about 200 cm. The main objectives in draining are to reduce the water content of the soil so that aeration will be adequate and so that dissolved salts will be removed and their movement towards the surface restricted.

The greater depth of drainage for irrigated land gives a better insurance against upward movement of salt and its accumulation near the surface. However, irrigated pastures and crops

which wholly cover the surface can be maintained with shallower drainage than that more usually required for irrigated orchards in which part of the ground is bare of vegetation and is not wholly covered by water at each irrigation. Methods of draining irrigated lands have been discussed by Luthin (1957a). Drainage is commonly required in irrigation areas and the need for it is usually to protect against salinity. An extended discussion of the movement of soluble salts is given in Section 8.

In relatively humid areas, the depth of drainage is sometimes rather carefully controlled so that some use may be made of the ground water when the rainfall during the growing season is not sufficient or is so irregular in its incidence that drought periods occur. Nicholson (1949) states that in such cases the optimum depends a good deal on the type of crops being grown (cereals required a lower groundwater level than root crops) and on the seasonal conditions (higher levels better in droughty summers than in moist summers). Because of these variables and the variability of soils, there has been much controversy about the depth at which water tables should be maintained to suit local conditions. Nicholson and Firth (1958) have reported on the effect of depth of the groundwater in a fen peat soil upon crop yields and soil structure. Their results (shown in Fig. 1) generally favour depths of 30 to 35 inches although potatoes and

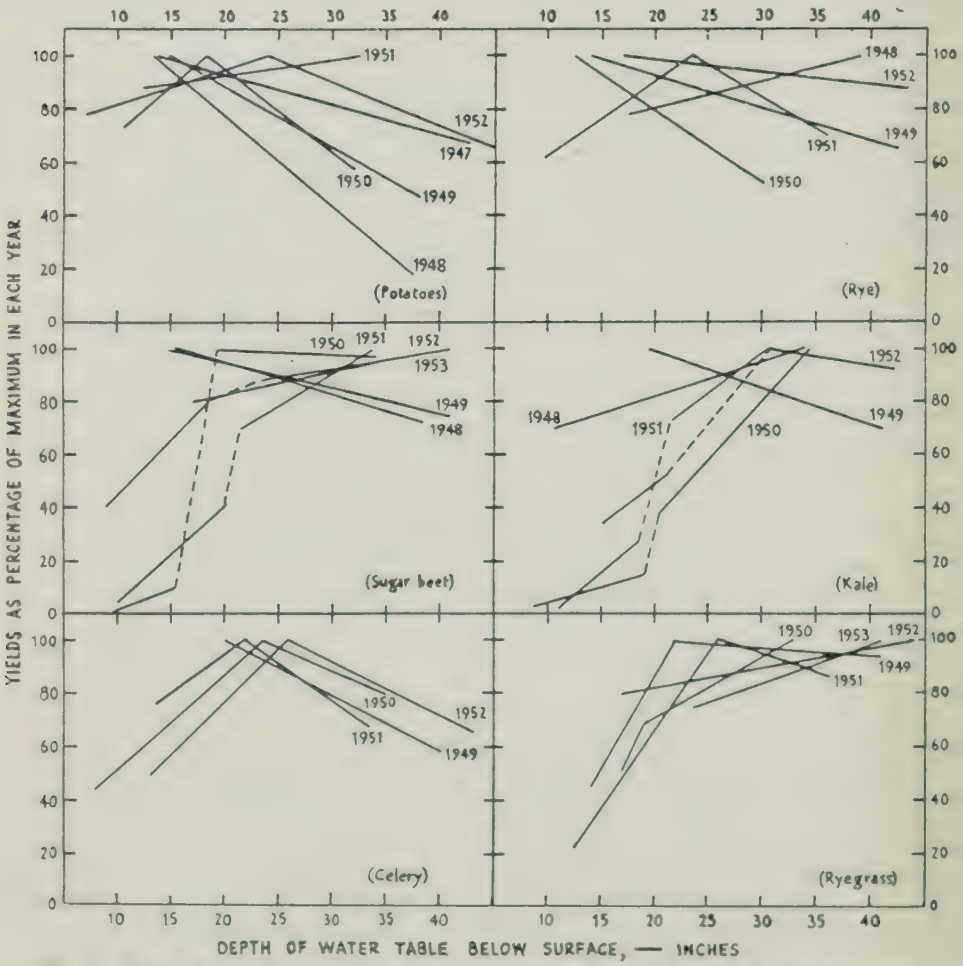

FIG. 1. *Effect of depth of water table on crop yields on a fen peat soil (Nicholson and Firth, 1958).*

celery responded to shallower groundwater levels (about 24 inches). There was a progressive deterioration in the physical condition when the water level was kept shallower than 30 inches during successive growing seasons. Reference can be made to Wessling and van Wijk (1957) and van't Woudt and Hagan (1957) for further details on depth of drainage in these circumstances and on sub-irrigation.

Wind (1955) has measured the contribution from a shallow water table to the root zone by inserting a plastic sheet above the water table. The water content of the soil above the sheet was lower throughout the summer in the soil treated in this way than in untreated soil and Wind found that 15 cm of water was supplied to the untreated soil during the season. For his clay soil, of measured unsaturated permeability, he estimated the relation shown in Fig. 2 for the amount of water (mm/day) contributed to the bottom of the rooting zone in relation to its height above the water table and the suction of water at that height. For example, if the suction is 1 atmosphere at a height of 40 cm above the water table, 3 mm of water per day will be contributed. The accuracy of these results is no doubt limited by the instruments used for measuring water content and suction in the field. Verhofen (1953) as quoted by Wesseling and van Wijk (1957), found a lower rate of supply of 3 cm per year in a soil column. Because rates of supply from below are not great, it is usually necessary under conditions of low rainfall and high transpiration to raise the water table periodically so that the soil will be properly sub-irrigated.

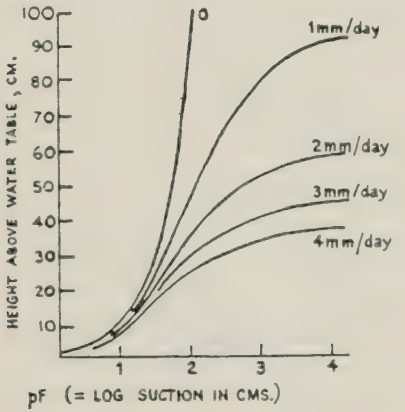

FIG. 2. *The amount of water (in mm/day) contributed by a water table to soil at a given suction and at a given height above the water table (Wind, 1955).*

Loss of production from having a water table too high (so that damage results when there is an unexpected rainy period) can be greater than from having it too low. For adequate aeration, Wesseling and van Wijk (1957) suggest that the fraction, ε_g, of the soil filled with air should exceed 0.1 and drainage should be designed to maintain conditions better than this for most of the time. The way in which diffusion of gases decreases with decreasing air-filled porosity (as caused by increasing water content) is illustrated in Fig. 3 by data of Taylor (1949). It will be seen that when ε_g falls to 0.1, diffusion is greatly restricted. In naturally occurring soil, non-uniformity would usually result in the sealing off of the soil at various planes and aeration could thus be stopped altogether at this stage. There is evidence of this in the literature especially in field measurements of diffusion (Blake and Page, 1948; Millington, 1957). Aslyng (1957) suggests that in tilling soil an aim should be to maintain ε_g at 0.15 or more but without making the soil so open that undue evaporation will occur.

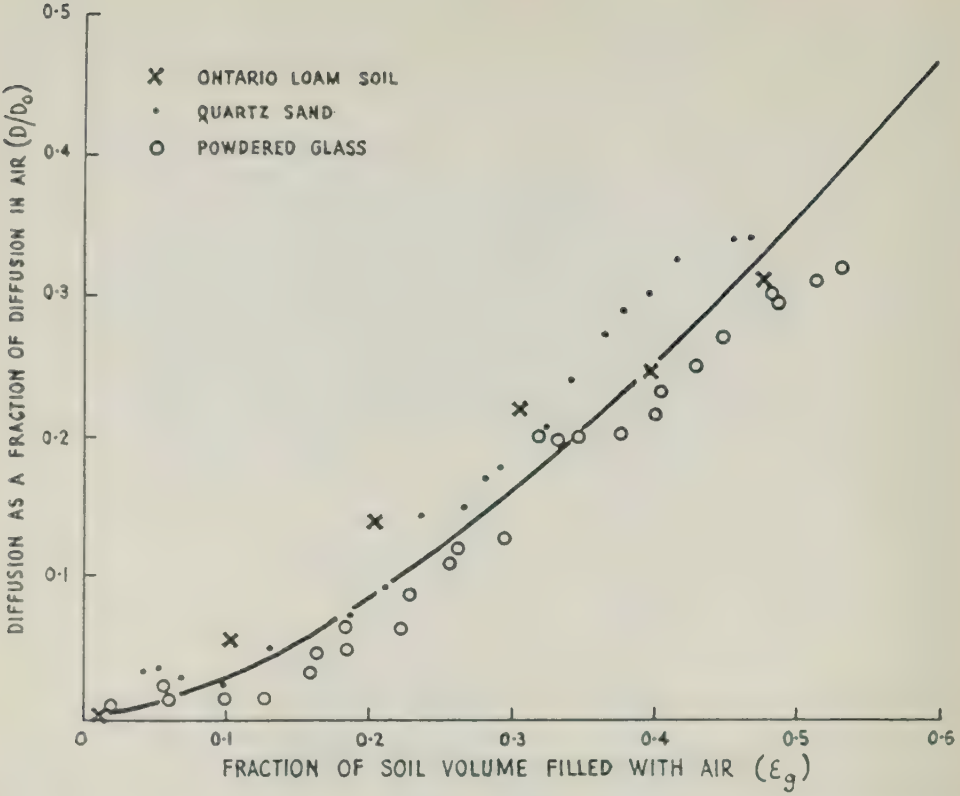

FIG. 3. *The effect of air-filled pore space on gaseous diffusion in three materials at various water contents (Taylor, 1949). The curve for Equation (27) in Section 7(II) is also shown.*

2. RETENTION OF WATER BY SOIL

1. Wetness of Soil

To find out how much water there is in a soil, a weighed sample is dried at some chosen temperature (usually 105°C) and the mass of water per unit mass of dry soil determined. This gives the water content, c_w in grams per gram, on a weight basis. For many purposes water content is better expressed on a volume basis. If the apparent (or bulk) density of the soil is ρ_a in g/cm³, then the volume of water per unit volume of soil is given by $c = c_w \rho_a/\rho$ cm³/cm³ and ρ, the density of water, may be taken as unity. Here the water content, c, represents also the equivalent depth of water in unit depth of soil. This is the quantity that is required when dealing with storage of water in soil and the gains or losses as by rain, irrigation, evaporation, transpiration or drainage.

These measurements of the quantity of water in soil do not give all the information needed for knowing how wet it is. A soil of high clay content can feel dry to the touch while a sandy soil at the same water content is quite moist. The first soil will not support plant growth while the second is wet enough to do so. Similarly water will tend to move from the sandy soil to the clay soil if they are in close contact although they are at the same water content.

This difficulty can be met in a rough way by expressing the water content as a ratio to some other value which is characteristic of the retentiveness of the soil for water. For example, if water content is expressed as a ratio to moisture equivalent (the water retained after centrifuging in a specific way) then the relative wetness of a soil is obtained. Other reference values which are characteristic of the colloidal properties of soil such as lower plastic limit, sticky point, or clay content have also been used instead of moisture equivalent in the same way.

A sounder and more direct way of getting this information is to find how strongly the water is held in the soil. This can be done by means of a tensiometer—an instrument for measuring the suction exerted by soil on the water enclosed in a porous ceramic cup to which a vacuum gauge or manometer is attached. The drier the soil the greater is the suction.

The wetness of a soil can therefore be specified in two ways—by the amount of water it contains and by the suction. The first specification is needed when working on the balance of gains and losses of water and the second is needed for an understanding of movement of water and availability of water to plants. The curve connecting amount with suction helps also to give information about the porous structure of a soil because water is withdrawn more easily from large pores than from small.

11. Suction of Soil

In Fig. 4, a method is illustrated for measuring suction. Porous cups attached to glass or plastic tubing containing water are inserted in the soil and water flows into or out of the soil through the cups until equilibrium is reached between the soil water and the unbalanced length, h, of the water column in the manometer.

The hydrostatic pressure of water, P, at any point in a body of water at rest (as in the manometer) is given by

$$P = P_o - \rho g h \text{ dynes/cm}^2$$

where ρ is the density of water, g is the acceleration due to gravity, and h is the vertical distance of the point from a reference level at which the pressure is P_o. If the reference level is the free surface of the water at atmospheric pressure where the reference pressure P_o is taken as 0, then

$$P = -\rho g h \text{ dynes/cm}^2$$

The negative sign indicates that pressure decreases with increasing height in a column of water, h having positive values when measured upward and negative when measured downward from the reference level according to convention. It can be seen that the pressure of the water in the porous cup of Fig. 4B is less than atmospheric pressure. The corresponding deficiency in pressure of water in the soil (taking this to be a sand for the time being) is given by the curvature

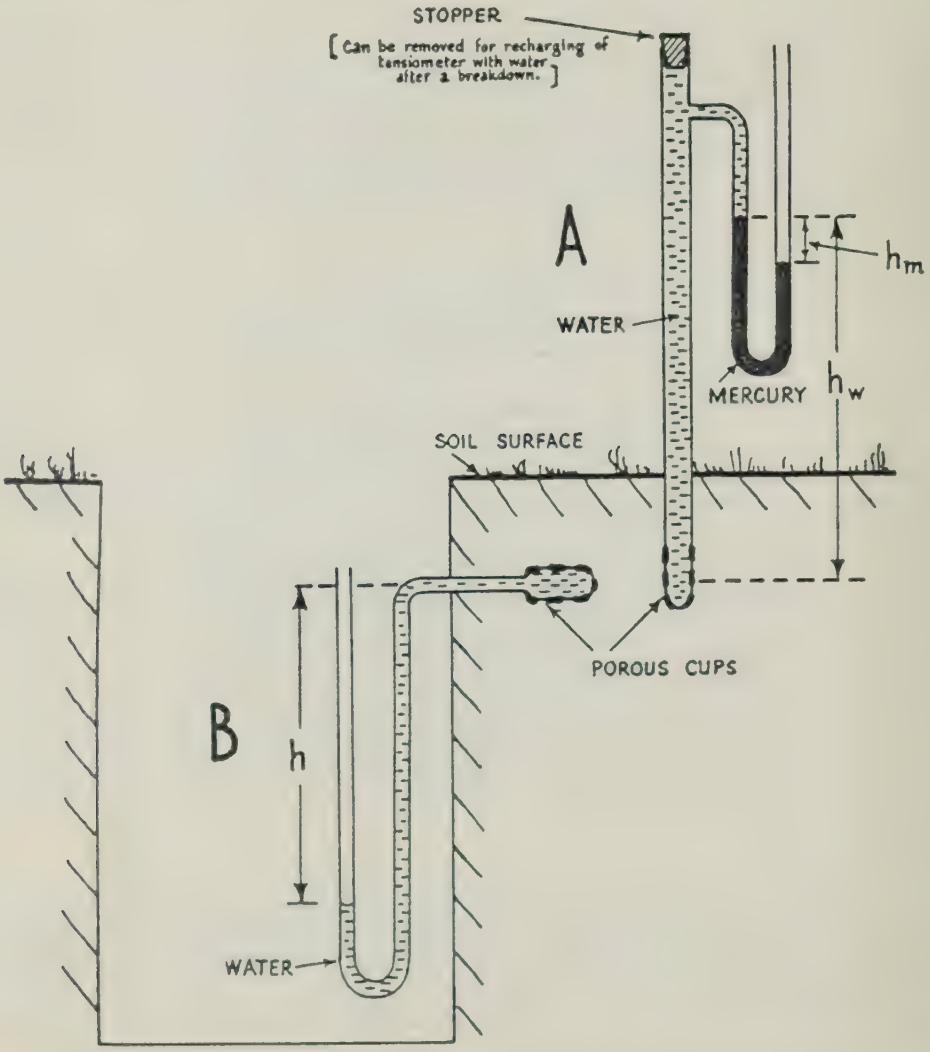

FIG. 4. *A.—Tensiometer for use in the field with a mercury manometer. [B.—A simple tensiometer in which the suction is shown as the length h of a column of water. This type of system is used in the laboratory for small suctions but is not ordinarily suitable for the field*

of air-water interfaces which exist within the unsaturated soil. When the air in the soil is at atmospheric pressure, the pressure, P, in the water on the other side of the interface is given * by

$$P = -2\sigma/r = -\rho gh \text{ dynes/cm}^2 \qquad \text{- - - - - - - - - (1)}$$

where σ is the surface tension of water and r is the radius of curvature of a simple hemispherical interface which would be in equilibrium with the more complex air-water interfaces of the porous system. For a non-circular pore, $1/r_1 + 1/r_2$ takes the place of $2/r$ where r_1 and r_2 are the principal radii of curvature of the interface.

To avoid the use of the negative sign in dealing with unsaturated soils, the water is said to be at a negative pressure, pressure deficiency, tension, or suction of ρgh dynes/cm^2. The dyne/cm^2 is not convenient as a working unit for suction which is consequently either expressed as cm of water (h) or as atmospheres (h/1035) for soils of low or high suction respectively. The bar ($=10^6$ dynes/cm^2) has also come into use as an alternative unit for suction (Richards, Gardner, and Ogata, 1956). It is close to 1 atmosphere (1 bar = 0.99 atmosphere) and the millibar ($= 10^3$ dynes/cm^2) is also close to 1 cm of water (1 mb = h/1.022). The bar and millibar are in general use as pressure units in other fields, especially meteorology. They are convenient for expressing both high or low suction in a consistent way. Inter-relations between these and other units for expressing water retention by soils are set out in Table 1.

Pressure deficiency constitutes a force capable of sucking water (at atmospheric pressure in bulk) into unsaturated soil. This is not however the only mechanism responsible for suction. Water can be drawn into a two-phase system of moist clay although there is no air-water interface within the soil mass. Here water is attracted to the solid surface and to the exchangeable cations associated with that surface. The range of operation of the attractive forces is limited and they are therefore of most consequence in the thinnest water films. In soils with very small particles (soils of high clay content) these thin films may together constitute a considerable amount of water because of the large total surface area of the particles and the ability of the soil to swell. Water is probably under a positive pressure near the solid surface (the pressure which can be developed by a swelling clay is evidence of this). Hence it may not be valid to speak of a negative pressure or pressure deficiency in soil water and for this reason the word "suction" is commonly used instead to cover the general case. Suction can be due to either mechanism and it is represented by the pressure deficiency of water in the porous cup of the tensiometer at equilibrium. In soils of low clay content, water retention can be treated as a capillary phenomenon (involving pressure deficiency) over the range of suction of main interest in agriculture. In soils of high clay content there is usually some filling and emptying of pores as well as change in volume upon wetting and drying and both capillary and attractive forces may be involved.

A number of different methods are used for measuring soil suction. The tensiometer is suitable for use when the suction does not exceed about 850 cm. As it increases beyond this, entrapped or dissolved air causes breakdown. Fig. 4A shows the general features of a tensiometer for measuring suction in the field using mercury as the manometer liquid. If ρ_m is the density of mercury, then suction is given by $s = \rho_m g h_m - \rho g h_w$ dynes/cm^2 $= h_m \rho_m/\rho - h_w$ cm water. Alternatively Bourdon vacuum gauges may be used. Details of tensiometer systems and similar laboratory devices (suction plates and tables) for measuring the relation between water content and suction have been reviewed by Richards (1949) who has played a leading part in the development of these instruments. The tensiometer cups are constructed with pores small enough not to let air be drawn through when the water in them is under a suction of about one atmosphere. Modifications to the usual type of tensiometer to enable suction to be measured without serious exchange of water between soil and tensiometer have been introduced by Miller (1951) and Croney (1952).

Richards (1941) has developed an instrument known as the pressure membrane apparatus by which the pressure difference across the interface can be measured for soil water at much greater suctions than are covered by the tensiometer. It consists of a closed chamber the floor

*—Assuming the contact angle between water and soil to be zero. The derivations and assumptions for equations used in this section are to be found in standard works on the properties of matter or in Edlefsen and Anderson (1943).

TABLE 1

Relation between the various units used for expressing retention of water (illustrated by approximate values at four levels of soil moisture)

Soil moisture condition	Suction		Capillary potential or free energy	pF	Equivalent radius of largest pores which could be full of water	Instruments suitable for measuring the suction
	length, h in cm, of vertical column of water or millibars	atmospheres or bars	ergs/g or ergs/cc	log h	10^{-4}cm or μ	
At suction of 1 cm (soil saturated or nearly so)	1	0.001	-9.8×10^{2}	0	1,500	tensiometer (or suction or pressure plate)
At suction of 100 cm (corresponding uncritically to field capacity)	100	0.1	-9.8×10^{4}	2.0	15	tensiometer (or suction or pressure plate)
At suction of 15 atmospheres (corresponding to permanent wilting point)	15,000	15	-1.5×10^{7}	4.2	0.1	pressure membrane apparatus
At relative vapour pressure of 0.85 (soil feels dry)	220,000	220	-2.2×10^{8}	5.4	0.007	vapour adsorption methods
Conversion from suction in cm (at 20°C when appropriate)	h cm or h/1.022 mb	h/1035 atm or h/1022 bars	$-980\ h$	log h	0.15/h	

of which is a membrane of cellophane or sausage casing suitably supported to withstand high pressures from inside. Water and dissolved salts can pass through this membrane (Reitemeyer and Richards, 1944) but it will hold back air when wet. Wet soil is placed on the membrane, compressed air or nitrogen is introduced into the chamber and brought to a chosen pressure. Water passes out of the soil through the membrane and continues to do so until the forces holding the water in the soil just balance the gas pressure tending to force water out of it. When flow ceases the soil water (under the influence of the gas pressure in the chamber) is in equilibrium with water outside the membrane at atmospheric pressure. If the gas in the chamber is then returned to atmospheric pressure, a suction of the same magnitude as the original gas pressure would need to be applied to the water outside the membrane to prevent it reentering the soil. Thus the pressure to which the gas was raised in the chamber is equivalent to the suction of the soil at the equilibrium water content. Pressures ranging from 0.1 to 170 atmospheres have been applied in this way although the working range for routine use does not commonly exceed 15 atmospheres. The water content of soil at this pressure ("fifteen atmosphere percentage" or "fifteen bar percentage" of Richards) corresponds rather closely to the permanent wilting point of soils.

Richards (1947) has published full details and drawings of the pressure membrane apparatus. A modification introduced later by Tanner and Hanks (1952) enables data to be obtained for the wetting (intake) curve as well as the drying (release) curve. This is accomplished by running water continuously below the membrane. The only serious criticism of the pressure membrane apparatus is that raised by Collis-George (1952) who found a progressive decline in water content of soil held under a constant gas pressure for several weeks. The decline was considered to be due to diffusion of gas through the membrane and the consequent necessity to introduce dry gas periodically into the chamber in order to maintain the pressure. However, according to Quirk (1953) this is probably not a major factor affecting the routine use of the apparatus.

The tensiometer (and similar suction devices) together with the pressure membrane apparatus (and the similar pressure plate apparatus of S. J. Richards (1938) and L. A. Richards and Fireman, 1943), enable the relation between water content and suction to be determined for a soil sample from saturation to permanent wilting point. For drier conditions than this, the vapour pressure of soil water is used as a measure of water retention (Bodman and Edlefsen, 1934; Schofield, 1938). It can be shown by means of the Kelvin equation that the suction in cm, h, is related to the relative vapour pressure, p/p_o, by the expression

$$h = -(RT/g) \ln (p/p_o) \qquad \text{- - - - - - - - - - (2)}$$

where R is the gas constant per gram, T is the absolute temperature, g is the acceleration due to gravity, p is the vapour pressure of the water in the soil and p_o is the vapour pressure of water in bulk. Since p will be affected by solutes in the soil solution as well as by the suction due to the solid matrix, Equation (2) will represent suction of the matrix only in the absence of solutes. It will be seen by reference to Table 1 that suction can reach very high values in the range covered by vapour pressure methods. For convenience suction is often expressed as $\log_{10} h$ for which the term pF was introduced by Schofield (1935).

In one of the many methods available for determining the relation between water content and vapour pressure, the soil is placed in a shallow container in an evacuated desiccator containing a saturated salt solution of known vapour pressure. The water content of the soil at equilibrium is determined for any chosen vapour pressure. Tables of vapour pressure for various saturated solutions have been published by O'Brien (1948). The vapour adsorption methods cannot at present be applied successfully to wet soil because there is very little difference in the vapour pressure of soil water from permanent wilting point to saturation. The relative vapour pressure changes only from 0.99 to 1 within this range. Vapour adsorption is used principally for conditions drier than the permanent wilting point but with refinement of methods (Steiger, 1951; Wylie, 1957) the range can be extended. There is therefore some promise of increasing usefulness of vapour pressure methods in soil physics.

Curves such as those in Fig. 11 and Fig. 6 which relate water content to suction are much used in soil water investigations. They have been given special names by some workers such as retention curves or moisture characteristics. An indication of the order of magnitude of the

suction under different moisture conditions and the relation between different methods for expressing water retention are given in Table 1. The measuring methods appropriate to each moisture condition are also given.

III. Suction and Pore Size

Curves relating water content and suction can also be used to help define the structure of soil in terms of the size of the pores which make up its pore space.
Equation (1) can be rewritten in the form

$$r = 2\sigma/\rho g h \qquad \text{- - - - - - - - - - (3)}$$

Here r is the upper limiting radius of pores which can remain full of water when a suction of h cm is applied to the water in wet soil. At 20°C, r is equal to $0.15/h$. The volume of water withdrawn in changing from one suction to a greater one in a soil which does not shrink, represents the volume occupied by pores of sizes lying between the corresponding limiting radii assuming the pores to be of circular section. The pore size distribution can therefore be determined directly from the water content—suction curve for a porous material as in Fig. 11. In this diagram, the curve represents the cumulative volume of pores smaller than a given size in a manner similar to that for representing particle size analysis.

The size distribution of the pores in a soil provides a specification of its structure which has been used in studying water and air movement. For example, permeability of saturated soil is related to the amount of larger pores and, in unsaturated soil, permeability is similarly related to the size distribution of those pores which are small enough to be full of water at the prevailing suction. Sometimes instead of using a complete curve the amount of pore space larger than one chosen size is determined ("non-capillary porosity" or "macroporosity"). A radius of 1.5×10^{-2} mm corresponding to 100 cm suction is frequently chosen arbitrarily for separating "large" from "small" pores.

IV. Some Limitations in the Use of Suction Data

There is one important difficulty involved in using a water content—suction curve. The curve exhibits hysteresis—it differs according to whether the soil is being wetted or dried. If two samples of a soil are brought to equilibrium with a given suction, a sample which was dry to start with will contain less water than another which was wet. Because soil is exposed most of the time to drying conditions when crops are being grown on it, the drying curve alone is commonly used. The question of which curve to select for measuring size distribution of pores is a more difficult one.

This hysteresis phenomenon in soils was early studied by Haines (1930). A number of different explanations have been advanced for it based on (1) possible differences of contact angle of water to soil as between wetting (advancing) and drying (receding), (2) possible effects of thickening of films before a hemispherical meniscus forms at the opening of the pore, and (3) the so-called "blocked pore" or "ink-bottle" effect. Hysteresis in soils is usually attributed to this last effect which is presumed to arise because many pores are larger than their openings. On wetting, these pores will not fill with water until the suction has fallen to that appropriate to the wide dimension of the pore proper. On the other hand they will not empty on drying until a higher suction has been reached appropriate to the narrow dimension of the neck. Hysteresis occurs in saturated clays for a rather different reason that will be discussed in Section 3 (II).

Carman (1953) prefers the drying curve for pore size measurements on porous materials because he considers that one or more connecting channels will usually be wide enough to allow a meniscus to enter easily into the pore. He attributes hysteresis more to delay in filling pores than to delay in the emptying of "blocked pores" during drying. This is the course usually followed by soil workers. Using drying curves (progressively increasing suction), Swanson and Peterson (1942) compared their results with size distributions measured microscopically. Although there was great irregularity in the shape of the pores, they found a close relation between results obtained by the two methods. Further comparisons of this sort are required.

The calculation of size distribution of pores from suction data depends on the conditions that loss or gain of water occurs without change in volume of the soil mass and that suction is dependent on the curvature of the air-water interface and not on adsorptive forces. These conditions exclude from consideration any clay soils of massive structure where swelling is important. To avoid changes in the geometry of such soils, liquids other than water can be used but little work of this sort has been attempted.

The use of vapour sorption data for calculating size distribution depends also on these conditions even although macroscopic swelling of the soil mass may not be observable. Under the relatively dry conditions covered by such data, the attractive forces at the surface of particles could be much more important than the film curvature. However, Carman (1953) and Pierce and Smith (1953) consider that the pore radius of many porous materials can be calculated from sorption data down to 20 Å which for water as the adsorbed liquid is equivalent to a relative vapour pressure (p/p_o) of 0.3 at 25°C. In soils, cation hydration may affect this picture but Schofield (1938) considers that r may be calculated for soils when $p/p_o > 0.5$. Schofield deduced this partly from the limiting occurrences of hysteresis in sorption isotherms. Zuev and Gapon (1943) reached a similar conclusion when they examined the relationships shown between the adsorption isotherms for Na, H, and Ca-bentonite. They found their series for increasing amounts adsorbed was $Na^+ < H^+ < Ca^{++}$ at low vapour pressures. But the curves started to come together in the range of p/p_o from 0.3 to 0.5 and they crossed over at 0.85. They considered hydration of cations was the main factor at low values of p/p_o, but in the middle pressures there was a transition and above 0.85 (corresponding to pores exceeding about 60 Å in radius) adsorption was due to condensation in the pores. However their evidence is not conclusive because as noted by Thomas (1928b) this effect of cations on adsorption above 0.85 could be attributed to swelling.

Little direct evidence has been advanced for the effect of soil structure upon adsorption of water vapour. The early work of Thomas (1928a) showed that soil structure affected the amount adsorbed when $p/p_o > 0.85$. A puddling pretreatment designed to increase the proportion of small pores increased adsorption while a freezing and thawing pretreatment which has a reverse effect on structure resulted in a decrease in adsorption. Carman and Raal (1951) have carried out a more comprehensive experiment in which they compared the adsorption of freon (CF_2Cl_2) on silica powder both as a loose powder and as a compressed plug. At low vapour pressures there was no effect of pore size, but at middle pressures more was adsorbed in the plug than in the powder because more small pores of the appropriate size were present in the plug. At high pressures near the saturation vapour pressure, the plug isotherm levelled off due to filling of the pores, and the powder isotherm continued to a higher level because more large pores were present in the powder.

From the rather tenuous evidence available, it appears then that the geometry of the porous system could be the dominating factor in water retention in soils when p/p_o is as low as 0.85. This is equivalent to a negative hydrostatic pressure or tension of about 220 atmospheres (Table 1) and the question arises as to whether the tensile strength of water is sufficient to maintain a tension as great as this. Temperley (1947) takes 500 to 1000 atmospheres as the order of the tensile strength of water on theoretical grounds. Experimental data are erratic but Temperley concludes from his own work and that of others that there is evidence for a strength of 60 atmospheres for water in a glass tube. More recently Briggs (1949) has reported a measured strength of 223 atmospheres. Even with some air dissolved in the water, strengths of the order of 40 atmospheres can be maintained according to Temperley and Chambers (1946) but more air (approaching saturation) reduced the strength to 6 atmospheres. Hence although tensions of some hundreds of atmospheres may theoretically be maintained in water, evidence for this in soil water in the presence of air is not yet available. This uncertainty does not affect other uses for suction data when suction is not taken specifically to represent negative pressure or tension.

Workers concerned with porous rocks have made much use of a mercury injection method for measuring size distribution in which mercury is forced into the pores. Scheidegger (1957) has described and illustrated apparatus used for this purpose. Because of the high surface tension of mercury, high pressures are required and this makes the method difficult to use in porous materials of low strength. Scheidegger has discussed the uses and limitations of the mercury injection and suction methods for measuring the size distribution of pores.

v. Solutes and Water Retention

Water retention has so far been discussed solely in relation to the porous matrix in which the water is embedded. A further way in which water is held has also to be considered when salts are present in the soil water. Solutes increase the osmotic pressure and lower the vapour pressure of water. They thus oppose water withdrawal from the soil through a semi-permeable membrane, through the roots of plants, or by evaporation. The total suction opposing water withdrawal in these cases is therefore made up to two components—a matrix suction (as already discussed) and a solute suction which is ordinarily referred to as the osmotic pressure of the soil solution. Solute suction can be determined from the properties of the extracted solution. The depression of freezing point of extracts from the pressure membrane apparatus is used by Richards and Campbell (1948) for this purpose.

Matrix suction is measured directly using the tensiometer, suction or pressure plate, and pressure membrane apparatus. Vapour pressure methods measure the total suction (matrix plus solute suction) directly.

The term solute suction has been proposed by Dr. H. E. Hayward and Dr. L. A. Richards of the U.S. Salinity Laboratory, Riverside, California, in place of osmotic pressure in dealing with water retention in soils. A complementary term referring to the solid framework would serve to distinguish the two components of total suction; and the term matrix suction has been adopted here. The word matrix was suggested to the writer by Mr. C. G. Gurr of the Division of Soils, C.S.I.R.O., Adelaide. Solute, matrix, and total suction are not terms of general currency and they have therefore been used only sparingly here. When the word suction is used alone, matrix suction is intended. Some of the terms currently used have been reviewed recently by the International Committee for Horticultural Congresses.

vi. Energy Relations of Water in Soil

If water is to be withdrawn from unsaturated soil, work has to be done against the hydrostatic pressure and adsorptive force fields or, conversely, a negative amount of work has to be done in adding free water to the soil. The amount of work, ψ, done against these fields of force in moving unit mass of water from a free water surface to the given point in the soil is given by

$$\psi = -gh \text{ ergs/g} \qquad \text{- - - - - - - - - - (4)}$$

$$\text{or} \quad \psi_v = -\rho gh \text{ ergs/cm}^3 \qquad \text{- - - - - - - - - - (5)}$$

where h is the suction or equivalent deficiency in hydrostatic pressure in cm of water and g is the acceleration due to gravity. The potential, ψ, is commonly referred to as capillary potential using the name originally introduced by Buckingham (1907). This name serves to distinguish it from the gravitational potential, $\varphi = gz$ ergs/g, which depends upon the height, z, of the given point above a chosen reference level. The rate of movement of water in the liquid phase is proportional to the gradient of $(\psi + \varphi)$. These potentials will be referred to again in dealing with water movement. They can be expressed either per unit mass or per unit volume of water and the subscript v is used here to indicate the volume basis. It is omitted in later sections where potentials are dealt with consistently on a volume basis.

The energy relationships can be considered also from the thermodynamic viewpoint as the change in free energy of unit mass of water in passing from free pure water to water at a given point in the soil. At this point the water may be at a different hydrostatic pressure under a curved air-water interface, it may be under the influence of adsorptive forces, or it may contain dissolved material. These will cause a lowering of the vapour pressure of the water in the soil and the resulting change in free energy, ΔF, per unit mass of water is then given by

$$\Delta F = RT \ln (p/p_o) \text{ ergs/g} \qquad \text{- - - - - - - - - - (6)}$$

where p is the vapour pressure of the soil water, p_o is the vapour pressure of free water, R is the gas constant per gram and T is the absolute temperature. In the absence of solutes the free energy can be set equal to $-gh$ ergs/g as in the case of capillary potential. The term pF which has already been referred to as standing for $\log_{10} h$ was introduced by Schofield (1935) on the basis that F indicated the connection to free energy of the soil water.

3. WATER IN SOIL OF HIGH CLAY CONTENT

I. Shrinkage Stages

In soil of high clay content, the geometry of the porous system can change with changing water content. Loss of water from swollen clay is accompanied by a reduction in the size of pores so that a clay block, lacking large pores to start with, may remain virtually saturated while shrinking even when dried to about pF 5.

Haines (1923), working with moulded blocks, noted that when a wet clay is dried, its loss of volume is at first equal to the volume of water withdrawn ("normal" shrinkage). Then a stage is reached at which air enters the block and the loss of volume is less than the loss of water ("residual" shrinkage). Finally no further change of volume occurs with loss of water. More recently, Lauritzen and Stewart (1941), Lauritzen (1948), and Stirk (1954) have extended this work to include shrinkage of natural clods and aggregates of high clay content. They found on drying from high water content that the loss of soil volume was at first less than the loss of water. This shows that in natural soil, there are pores which allow air to enter as water is withdrawn. They consider this preliminary stage to be an important attribute of natural soil as distinct from moulded blocks since it assists aeration. It is a characteristic of porous aggregates, but is not well represented in massive subsoils. This shrinkage stage was termed "structural" by Stirk.

Schofield (1938) deduced from Haines' data that the transition from normal to residual shrinkage took place at about pF 5 in Haines' blocks; but until recently no direct measurements of shrinkage in relation to suction have been made. In a detailed study of suction—water content —volume relationships of blocks made from two clay soils, Holmes (1955) found that residual shrinkage commenced at pF 5.1 in one case and 5.5 in the other. Shrinkage ceased after pF 5.7 had been reached in the first soil but continued to pF 6 in the other. Stirk (1954) found that the limiting suction beyond which there was no further shrinkage of aggregates increased with increasing clay content of the aggregates.

II. Movement of Soil During Shrinking and Swelling

It is obvious from the crack pattern at the surface of a dry clay soil in the field that horizontal movements contribute to volume change. The extent to which vertical movements also contribute under field conditions is not so clear although it is believed that they can affect microrelief as in gilgai formation (Leeper, Nicholls, and Wadham 1936; Hallsworth, Robertson, and Gibbons, 1955). The settlement of buildings and roads (Ward, 1949; Croney and Lewis, 1949) following withdrawal of water by vegetation is another manifestation of vertical movement of soil due to change in water content.

Vertical movements of soil with a grass cover and subject to seasonal changes in water content, have been measured by Ward (1949) at Garston, England, and Aitchison and Holmes (1953) at Adelaide, Australia. Their magnitude is dependent on soil, climate and cover. The seasonal movements in a soil of high clay content resulting from water gains during wet cool winters and losses during dry hot summers are illustrated in Fig. 5 from the data of Aitchison and Holmes. Movement is greatest at the surface because the full effect of underlying horizons is to be observed there. Seasonal displacement of the surface was as much as 1.5 inches in this soil. Observations such as those of Fig. 5 are important in the design of foundations for masonry buildings which are liable to crack badly because of unequal movements of clay soils under the structure. In order to reduce the effects of vertical movements, footings of building foundations may be placed at sufficient depth in clay to avoid the effects of seasonal changes in water content.

Volume change of a small isolated block of clay is usually the result of equal changes in three dimensions (unless there is some preferred orientation of clay crystals in one plane). If this process of three-dimensional or isotropic volume change occurs in the field then cracks will

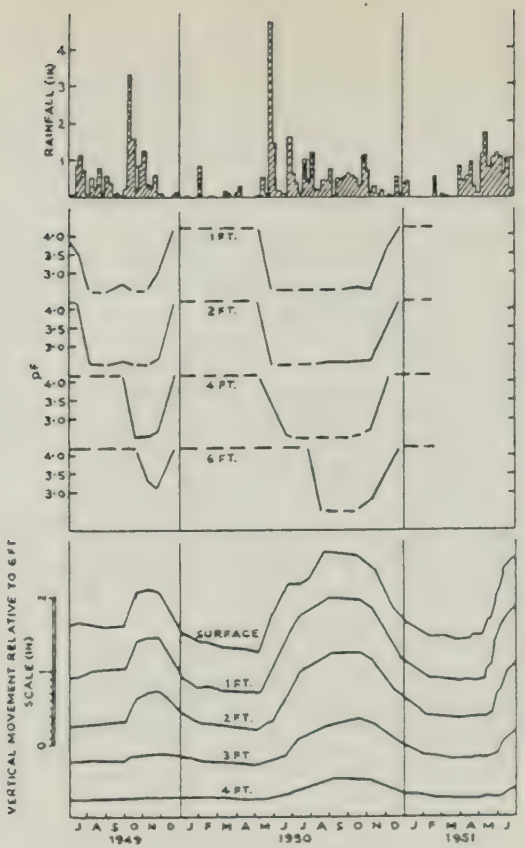

FIG. 5. *Vertical movements of a soil of high clay content accompanying seasonal changes in suction (Aitchison and Holmes*, 1953)

form on drying and close again on wetting and the soil will fall and rise with a minimum of rearrangement of the particles amongst themselves. But it is conceivable that, under field conditions, the load of overlying soil (overburden) which has to be supported could cause plastic flow in shrinking. If so, much of the volume change would be represented by vertical movement (one-dimensional or anisotropic change of volume) and cracks in subsoil would not be a major feature on drying. Aitchison and Holmes compared their vertical movements with corresponding gains or losses of water in the underlying soil. The results which they obtained on clay subsoils (Fig. 5 and other data) were compatible with isotropic, three dimensional, change assuming volume change to be "normal". Their conclusion applied to clays with an overburden which was mainly less than four feet.

Cracking resulting from shrinkage can help water entry, drainage, and air circulation very greatly in clay subsoils provided the soil has been dried out rather thoroughly (Stirk, 1954). The structural units, which are obvious when dry, possess a degree of permanence on wetting (Russell, 1950, p. 399). The swelling of clay is a slow process and it may be some months after a wet season commences before the cracks close up even when water is freely available. As a consequence the complete desiccation of a clay profile affects its behaviour to water and air for some time after it is rewet. Emerson (1955) has shown that the drying out of the subsoil by deep-rooted crops during the summer results in improved drainage in the following winter in England.

He found that the water content of isolated aggregates continued to increase over a period of three months at low suction. He considered that this was due to the rearrangement of the clay crystals within the aggregates, as the thickness of the water film around them gradually increased, and that this was accommodated by a gradual reduction in the size of the largest pores within the aggregate. Permeability would then fall off progressively. This mechanism is probably complementary to the slow closing of cracks outside the aggregates as they gradually increase in volume. The crack-closing mechanism is the more likely to occur in those clay subsoils where the aggregates are so dense as to be almost saturated even at suctions as high as pF 5.

The major movements of soil are certainly accompanied by small scale rearrangements of particles of the type considered by Emerson. In this connection Holmes (1955) studied the hysteresis loop in the wetting and drying curves for the water content-suction relation of clay blocks. The ordinary explanations of hysteresis (based on the presence of an air-water interface) were not tenable because his clay blocks showed hysteresis although saturated with water. Holmes, following an early suggestion by Schofield (1935), therefore attributed hysteresis in clay to the energy changes involved in readjustments of particles relative to their neighbours.

Such readjustments can result in the preferred orientation of clay crystals in one direction. Weymouth and Williamson (1953) have shown that extrusion or any shearing movement in moist clay may cause this. When it has occurred, orientation will affect shrinkage behaviour markedly. Williamson (1947) has shown that discs made of oriented clay (kaolinite plus some illite) had a relative shrinkage in the direction perpendicular to the basal plane which was more than twice that in a radial direction. Orientation must play some role in the structure and swelling of many soils; but little is known about this at present. Brewer and Blackmore (1956) have shown that aggregates of some soils have patches and coatings of oriented clay and they suggest that breakdown on wetting occurs because of these oriented zones. Swelling is greater in one direction than it is in others in these zones and thus stresses are set up when the soil is wetted rapidly. This is further discussed in relation to aggregate stability in Section 11.

III. Retention of Water in Clays

In sandy soils the water content at a given suction depends on how much pore space there is of the right size to hold water. In soils of high clay content, the filling and emptying of pores on wetting and drying may be less important than swelling and shrinking. The relation between water content and suction depends then less upon the pore structure of the soil and more upon the size distribution of the particles and their surface properties.

The exchangeable cations in a water layer occupying the space between two parallel clay plates give rise to a higher osmotic pressure than exists in bulk water free from exchangeable cations and hence of lower ionic concentration. Since the exchangeable ions are not free to migrate and so eliminate the concentration gradient, there is a tendency for more water to be drawn into the layer between the particles. If this tendency is not resisted by a confining pressure equal to the swelling pressure, the clay plates will be pushed further apart and the soil swells.

Methods for calculating the swelling pressure have been derived by Schofield (1946) and others. These depend upon a knowledge of the ionic distribution in the electrical double layer so that the concentration and hence the osmotic pressure in the plane midway between the clay plates can be calculated for a water layer of known thickness. The difference between the osmotic pressure of water in this plane and of bulk water or solution in equilibrium with it represents the swelling pressure. Bolt (1956) has discussed the principles and assumptions involved in the calculation of swelling pressure from double layer theory and has tested it against experimental data for the relation between water content and external pressure in a two-phase water-clay system. Bolt followed the consolidation of clay when pressure was applied and water and solutes were allowed to escape through a permeable membrane (a system similar in principle to the consolidometer in soil mechanics). In this experiment the extracted solution represents the bulk solution of known concentration. Agreement between experiment and calculation was good enough to lend support to double-layer theory as a basis for explaining swelling (and hence also water retention) in Na-montmorillonite. Work on Na–illite and Ca–montmorillonite was also done but only qualitative agreement between experimental and calculated data was obtained.

Swelling has been explained in other ways. Terzaghi (1931) considered it to be a mechanical phenomenon. The small plate-shaped particles bend elastically when pressed together by an external pressure or by tension in the water on drying. Upon removal of the applied pressure or upon wetting, restitution of the system proceeds and expansion occurs. Bolt has pointed out that there is no evidence for this type of mechanism in the swelling of clays but that, on the other hand, the application of double-layer theory is well supported by evidence that clay plates are charged and have exchangeable cations associated with them. However further work is needed to establish this application of the theory. Terzaghi and Peck (1948, p.15) and Terzaghi (1956) suggest that both mechanical and adsorptive mechanisms play a part.

Kryukov and Komarova (1954) have examined the effect of very high pressures of up to 20,000 kg/cm^2 (almost 20,000 atmospheres) on the expulsion of water from a layer of material 5 mm thick. They obtained approximately linear curves for the relation between water content and the logarithm of the applied pressure for bentonite and kaolin. Starch did not follow this relation and it retained more water than the clays at the highest pressures although it has little mechanical strength. They concluded that the retention of water in these materials is governed primarily by the physcial properties of the water itself conditioned by its interaction with the solid and not by the mechanical strength of the solid skeleton.

It is evident that water may be withdrawn from a swollen clay by applying a load to the clay or a suction to the water and that consolidation may be accomplished by either process (Terzaghi, 1943, p. 333; Croney and Coleman, 1953). The changes in volume brought about by changes in the pressure applied to soil in a consolidometer on the one hand and by changes in suction applied by means of suction plate, pressure membrane apparatus, and vapour sorption methods on the other have been compared by Aitchison and Donald (1955). In a non-swelling loam soil, the volume was unaffected by suction although affected by applied pressure. In a kaolin sample and in soils of high clay content the application of any suction up to about 25 atmospheres had approximately the same effect upon volume as compressing it at the same pressure in a consolidometer.

The effect of the overlying load upon suction in subsoil clays has been studied by Croney and Coleman (1953) who have shown that the suction, h, in cm of water in a loaded soil *in situ* will be

$$h = h_s - \alpha H \qquad \text{- - - - - - - - - - (7)}$$

where h_s is the suction of an "undisturbed" sample without load, H is the pressure of overburden expressed as height of an equivalent water column, and α is the fraction of the applied pressure affecting the suction. If the soil is non-swelling, $\alpha = 0$; if it is entirely colloidal, and saturated with water $\alpha = 1$. Values for α as determined by Croney and Coleman from the other quantities in this equation, varied between 0.1 and 0.6 at various depths in a silty clay. Calculations made independently from shrinkage tests confirmed these values. In saturated soil of high clay content it is reasonable to assume that $\alpha = 1$. In this case a soil of given water content, which has a suction of (say) 200 cm in an undisturbed sample, has a suction of zero when in place at a depth of 100 cm if soil of wet bulk density of 2 g/cc lies above it. Equation (7) shows how the effect increases with depth and is unimportant at high suctions. It has been used for calculating the moisture distribution above a water table under pavements. Russam (1958) has reported agreement with measured moisture distribution under roads in Trinidad.

The structure of soil of high clay content is affected by the maximum load or the maximum degree of desiccation to which it has been subjected in its history. In soil free from the influence of a permanent water table, the effect of desiccation will usually be the more important because of the great magnitude of maximum suction in comparison with overburden load at all ordinary soil depths. By means of the curves of Fig. 6, Croney and Coleman (1954) have shown how the swelling behaviour of clay is affected by the highest suction (driest condition) it has previously reached. Clay dried from a slurry (curve D) will not rewet to that condition. It will instead re-wet along a new, reproducible curve (E) with less water at a given suction than originally, and it can then be made to dry back so as to close its hysteresis loop at the same dry limit. If it is subsequently dried beyond that limit, a different reproducible wetting and drying cycle (F) is obtained with still less water at a given suction than before. The drying curve (D) for the slurried soil does not join the drying curve for the natural soil until pF 4.8 is reached because the natural soil has in its history been previously exposed to drying conditions or pressures of this magnitude.

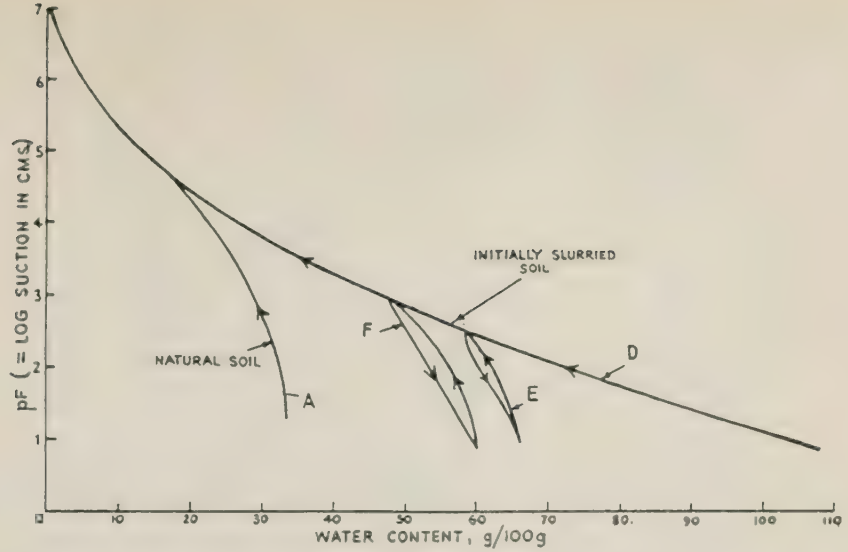

FIG. 6. *Relations between suction and water content for a clay soil* (*Croney and Coleman*, 1954)

IV. EFFECT OF SOIL PROPERTIES ON SORPTION OF WATER VAPOUR

The sorption of water vapour by soil has long been used as a means for studying soil-water relations in the relatively dry range. Thomas, in a series of papers between 1921 and 1928, showed that the amount of water adsorbed at a given relative vapour pressure increased with increasing clay content and depended also on the type of exchangeable cation and the salt content. Also, as already discussed, he recognized that condensation in pores played a part. He observed that the curve relating water content and relative vapour pressure (the sorption isotherm) was sometimes sigmoidal and this was shown by Puri, Crowther and Keen (1925) to be the general shape of the curve for soils.

The early work had qualitatively established the influence of amount and composition of clay particles and their exchangeable cations on sorption. Hence it is not surprising that most of the recent work has dealt with the clay fraction extracted from soils or, more particularly, with clay minerals from recognized deposits.

The discovery of the crystalline nature of clays made possible their identification and stimulated the study of their properties. It was shown that montmorillonite with its expanding lattice and small particle size adsorbed more water than kaolinite with its larger particles and that illite occupied an intermediate position. Typical curves are given in Fig. 7 for these clay minerals, all three of which are of common occurrence in soils. Hysteresis in the wetting and drying processes is characteristic of this as of other types of suction measurement and it is illustrated in Fig. 7 by the difference between the solid and the broken lines for wetting and drying conditions respectively.

Most of the available data on the effect of exchangeable cations on adsorption support the generalization (Jenny, 1932) that the amount of water adsorbed increases with the charge of the cation and decreases with its size. Rios and Vivaldi (1950) consider that the amount adsorbed at a given vapour pressure is a linear function of the charge of the exchangeable cation divided by the square of its radius and they show that data of their own and of Hendricks, Nelson and Alexander (1940) fit their equation.

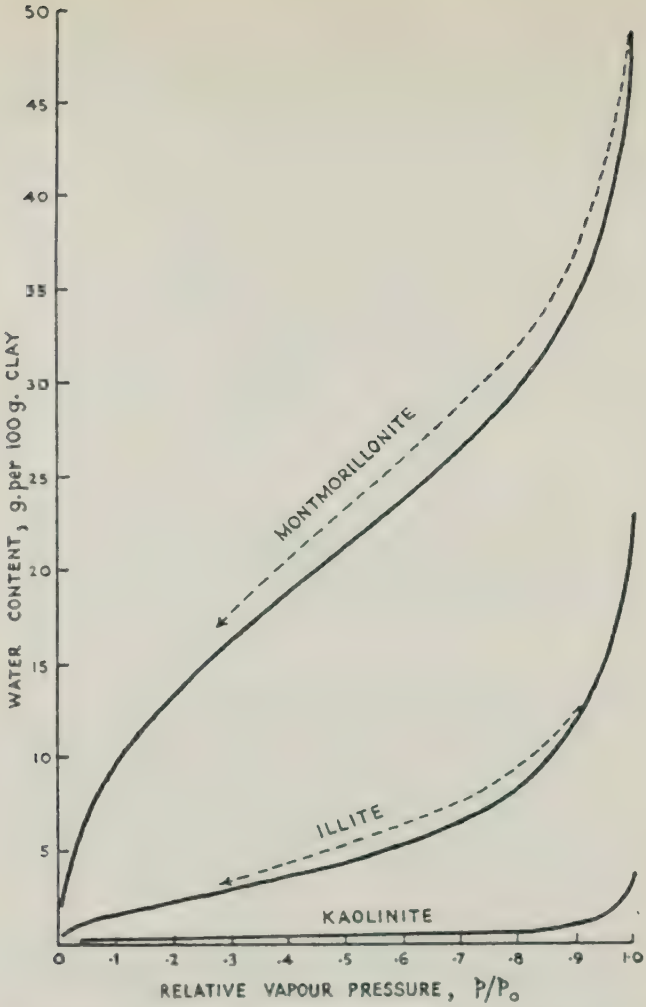

FIG. 7. *Sorption of water vapour by different clays. The continuous lines represent wetting curves for adsorption and the broken lines drying curves for desorption* (*Orchiston*, 1954).

The adsorption isotherms of Keenan, Mooney and Wood (1951) and Mooney, Keenan and Wood (1952b) provide sets of data showing the effects of cations on adsorption of water vapour by kaolinite and montmorillonite. With the aid of the multimolecular theory of adsorption of Brunauer, Emmett and Teller (1938) which will be discussed below, these workers have calculated the number of water molecules associated with each cation at a stage corresponding to the formation of the B.E.T. "monolayer". Their calculations are collected together in Table 2. These demonstrate the greater association of water molecules (hydration) with divalent than monovalent ions and the inverse relation between hydration and size of ions having the same charge. Hydrogen and lithium are exceptions to these general relations.

TABLE 2

Hydration of exchangeable cations (from Keenan et al, 1951, and Mooney et al, 1952b)

Cation	Radius Å	Molecules of water per cation	
		Kaolinite*	Montmorillonite+
H^+		3.5	3
Li^+	0.60	0	1
Na^+	0.95	2.4	
K^+	1.33	1.6	
Rb^+	1.48	1.5	
Cs^+	1.69	1.3	
Mg^{++}	0.65		8
Ca^{++}	0.99	10	8
Sr^{++}	1.13	6	4
Ba^{++}	1.35	5	4

*—Average values based on evidence of no hydration of lithium ions in kaolinite.

+—Assuming no hydration of rubidium and caesium ions for the purpose of calculation.

These hydration values summarize the relative effects of the exchangeable cations on adsorption of water by clays at low relative vapour pressures. In making comparisons, allowance has to be made for the fact that, in homoionic clay, there are two monovalent cations for each divalent cation. Also for the purpose of calculation the hydration of cations on montmorillonite was expressed relative to zero hydration of the largest monovalent cations, caesium and rubidium (which according to the kaolinite data are hydrated to the average extent of about 1.5 molecules). Neglecting the behaviour of lithium in kaolinite, the sequence of increasing adsorption is found then to follow rather closely the series Cs^+, Rb^+, K^+, Na^+, Li^+, Ba^{++}, Sr^{++}, Ca^{++}, Mg^{++}. Results with H-clay are subject to possible effects due to the release of aluminium at low pH. Most workers show adsorption by H-clay to be equal to or less than Ca-clay. In the present series hydrogen falls between calcium and strontium.

The detailed adsorption data on which Table 2 was based show that the sorption isotherm for montmorillonite is not strictly a smooth curve. A step-wise increase of water content occurred with increasing vapour pressure due to the crystalline swelling of montmorillonite which is known from X-ray information to progress in a step-wise fashion. Collis-George (1955) noted this also in Na-montmorillonite using pressure membrane apparatus as well as vapour adsorption; but not all workers (see Fig. 7) have observed steps in these adsorption isotherms.

Mering (1946) has compared the crystalline swelling of Na- and Ca-montmorillonite using vapour adsorption and X-ray methods. The swelling of Na-montmorillonite in relation to relative vapour pressure is illustrated in Table 3 by X-ray data and observed changes in volume of a flake.

TABLE 3

Swelling of an oriented flake of Na-montmorillonite (Mering, 1946)

Relative vapour pressure	Water adsorbed, g/g	Crystalline swelling Average spacing Å	Apparent swelling
0	0	9.6	No swelling
0.9	0.4	16.2	Beginning of swelling and plasticity
0.96	0.6	18.5	About 30% swelling
0.99	1.0	19.5	About 100% swelling

A large amount of water (40% by weight) is adsorbed by crystalline swelling without causing any apparent swelling of the flake up to a relative vapour pressure of 0.9 (or 90% relative humidity). This water is accommodated by an increase in the spacing between the elementary silicate sheets of the clay crystal from 9.6 to 16.2 Å. Apparent swelling then occurs and there is an accompanying increase in spacing to 19.5 Å at a relative vapour pressure of 0.99. Thereafter water uptake continues when the sample is placed in contact with water and apparent swelling proceeds up to a 20-fold increase in volume.

There is not full agreement on what happens to the clay crystal when wetting proceeds beyond 0.99 relative vapour pressure. Mering found no increase in spacing beyond 19.5 Å but there was increasing disorder of the elementary sheets. Collis-George found no evidence of expansion beyond that spacing but considered that the multiple sheets (crystals) persisted even in dilute suspensions. On the other hand, Norrish and Quirk (1954) found evidence of the persistence of crystalline expansion well beyond this limit.

Ca-montmorillonite undergoes stepwise crystalline swelling upon wetting until a maximum spacing of about 20 Å is reached. Upon further wetting of the flake the crystals remain stable and the sheets do not separate. In contrast with Na-montmorillonite, apparent swelling was restricted in Mering's oriented flakes of Ca-montmorillonite at relative vapour pressures greater than 0.9.

X-ray studies of expanding clay crystals enable critical measurements to be made of the distance between parallel sheets during swelling. This cannot be done directly for the disordered expansion of other clay minerals. Hence although montmorillonite is not to be regarded as the most common or typical clay mineral in soils, it is a useful material for studying swelling of platy clay particles in general. These studies of crystalline swelling, water sorption, and ion hydration have in the main been done at relative vapour pressures of less than 0.99. Under wetter conditions than that, swelling may be explained more generally (as in the preceding section) as being due to osmotic pressure resulting from the exchangeable cations in the diffuse double layer. Thus the present incomplete picture of swelling is based on the pushing apart of the elementary sheets and of the clay crystals by increasing the number of water molecules associated with the exchangeable cations.

v. Sorption and the Surface Area of Particles

By means of theories of physical adsorption (especially that of Brunauer, Emmett and Teller) attempts have been made to calculate the specific surface (surface area per gram of soil) from the adsorption isotherm for water vapour. An ordinary particle size analysis tells nothing about the size distribution of the colloidal particles some of which may be much larger than others although all fall within the clay fraction. It is therefore useful to know the surface area of soils and clay minerals (for this and other purposes too) and, if it could be calculated from such simple measurements, this would be an attractive step. Specific surface of soil particles can vary widely from about 100 cm^2/g in sandy soil to several hundred m^2/g in some clay minerals.

The calculation of surface area in this way is complicated by the effects of exchangeable cations on adsorption of water vapour. The assumption has to be made that, in the first stage of adsorption, the water molecules pack together in a regular fashion to form a unimolecular layer over the whole surface of the particles. There is no general agreement on the configuration of water molecules adsorbed on clay surfaces (Grim, 1953); but it is certain that a uniform packing area does not exist because of the effect of exchangeable cations. These provide active spots for adsorption of water and their effect varies with the particular cation.

The use which is being made of the theory of multimolecular adsorption of Brunauer, Emmett and Teller (1938) in studying adsorption of water vapour by soils and clays, warrants some discussion of the theory itself as developed for other, less complex, systems. The physical adsorption of a gas (the adsorbate) on the surface of a solid (the adsorbant) is an exothermic process and the heat of adsorption decreases for each successive layer of molecules covering the surface. With the aid of the simplifying assumption that the heat of adsorption of all layers but the first on the plane surface of a solid is equal to the heat of condensation of the adsorbate, Brunauer, Emmett and Teller (1938) developed their theory of multimolecular adsorption.

This is referred to as the B.E.T. theory. Their justification for the assumption lies in the short range of action of the forces supposed responsible for physical adsorption. The following equation which arises from their theory has been found by a number of workers to fit data for the adsorption of nitrogen or water vapour by soil and clay minerals,

$$\frac{p}{v(p_o - p)} = \frac{1}{v_m c} + \left(\frac{c-1}{v_m c}\right)\left(\frac{p}{p_o}\right) \qquad \text{- - - - - - - - - (8)}$$

where v = total volume of gas adsorbed

v_m = volume of gas adsorbed when the entire adsorbent surface is covered with a complete unimolecular layer

p = pressure of the gas

p_o = saturation pressure of the gas.

c = a constant involving the heat of adsorption in the first layer and the heat of condensation.

For a set of experimental data obeying the theory, a straight line is obtained when $p/v(p_o - p)$ is plotted against p/p_o. Since the slope of this line is $(c\text{-}1)/v_m c$ and its intercept is $1/v_m c$, the values of the constants v_m and c can be obtained from the plot. The number of molecules in the unimolecular layer can be obtained directly from v_m, and the surface area can be calculated on the assumption of a particular packing area per molecule of adsorbate. In the case of adsorption of water vapour on soil where adsorption is measured gravimetrically, the mass w and w_m can be used in place of the volume v and v_m respectively in Equation (8). The equation is fully discussed by Brunauer (1943).

Emmett, Brunauer and Love (1938) have measured the surface area of soils and soil colloids using nitrogen (at −183°C) so as to be free from any reaction between the gas and exchangeable cations. However a nitrogen adsorption isotherm does not always provide a measure of the total surface which will be exposed for the adsorption of water vapour. Mooney, Keenan and Wood (1952a) determined the adsorption of water vapour and of nitrogen by montmorillonite. Using a B.E.T. plot and assuming packing areas of 16.2Å² and 10.8Å² per molecule of nitrogen and water respectively, they found a monolayer area of 27 m²/g with nitrogen and 422 m²/g with water. From these two values they calculated the total area of 817 m²/g on the following basis. Nitrogen does not penetrate between the sheets of the crystal lattice of montmorillonite and hence covers the external surface only. Water enters between the sheets but at the stage of monolayer formation as derived from the B.E.T. plot, the one layer between the sheets is shared by two surfaces. Hence the total surface area is given by twice the area obtained using water minus the area obtained using nitrogen.

Goates and Hatch (1953) have applied the B.E.T. equation to the data of Alexander and Haring (1936) for the adsorption of water vapour on soil colloids (<3μ) from the Cecil, Miami, Carrington and Barnes soils in U.S.A. and, from this, the surface areas are found to be 70, 127, 184 and 303 m²/g respectively. They have been altered from the original paper of Goates and Hatch in which some errors are apparent. These surface areas were obtained from water data only and, as pointed out by the authors and as discussed above, they represent external surface plus half the internal surface of any expanding lattice minerals. The dominant minerals are probably respectively kaolinite, illite, and montmorillonite in these soil colloids.

Orchiston (1953) has calculated the surface area of New Zealand soils on which he had obtained data for adsorption of water vapour. In addition to the B.E.T. equation, he also used an equation developed by Harkins and Jura (1944). In his series of six soils, with surface areas ranging from 20 to 200 m²/g, estimates of surface area were close by the two methods. Similarly for the clay minerals whose isotherms are illustrated in Fig. 7, Orchiston (1954) found the following surface areas by the two methods.

TABLE 4

Specific surface area of natural clay minerals from adsorption of water vapour

Clay mineral	Specific surface area, m^2/g	
	From B.E.T equation	From equation of Harkins and Jura
Montmorillonite+	454	501
Illite	86	82
Kaolinite	6	10

+—See text for correction needed because water vapour is the adsorbate.

The B.E.T. equation has been more used with soils than the Harkins and Jura equation which does not always give a near linear plot. Wadsworth's (1944) data can be shown to fit the B.E.T equation but to give an unsatisfactory plot with the other, and Keenan, Mooney and Wood (1951) found this to be the case also for their own data on adsorption of water vapour by kaolinite.

The stage of wetting w_m, represented in the B.E.T. theory of adsorption by the completion of the first molecular layer, was found by Orchiston (1953) to correspond on the average to a relative vapour pressure of 0.21 in clays. Similarly the values for w_m given by Goates and Hatch can be shown from Alexander and Haring's original data to correspond to values of p/p_o ranging from 0.17 to 0.25. If 0.21 represents roughly the relative vapour pressure of the water when the B.E.T. monolayer is complete, then from Equation 6 in Section 2 it follows that the free energy of the soil water at that stage is in the order of -2.1×10^9 ergs/g or pF 6.3.

In most cases where the B.E.T. plot has been examined, a good fit to a straight line has been found for values of p/p_o between 0.05 and 0.35—the usual range of applicability of the theory—but these cases do not in themselves prove that the B.E.T. theory holds for the adsorption of water vapour by soil. Because of the effects of exchangeable cations on adsorption, the soil water system differs from those systems on which the adsorption theories were originally developed and checked. On this point, Quirk (1955) found poor agreement between values for surface area of clays determined by adsorption of water vapour and of nitrogen at low temperature. The difficulty, as discussed by Keenan *et al*, is that clays do not present a surface of uniform adsorption energy and that the packing area for water on clay varies with the exchangeable ion. These workers consider that the B.E.T. theory is nevertheless applicable because the only essential requirement of the theory is that the energy of adsorption of the first group of molecules should be different from the rest. But it does not follow that it enables surface area of clays to be measured when water vapour is used as the adsorbate.

In the field of soil physics the general trend has been to treat the relation between water content and suction as a continuous one without any critical breaks corresponding to such stages as field capacity, permanent wilting point, limit of capillary condensation, or adsorption of a first "layer" of water molecules. These stages are all more or less arbitrary and the values for p/p_o of (a) 0.85 and (b) 0.2 respectively can only represent orders of magnitude of the relative vapour pressure down to which (a) the pore size distribution is significant in water retention and (b) below which the first group of molecules associated with the absorbing surface is being removed. They cannot be regarded as critical limits.

4. PERMEABILITY OF SATURATED AND UNSATURATED SOIL

1. Introduction

If open tubes are inserted at two points in uniform saturated soil through which water is passing, the difference in the level of the water surface in the two tubes will give the difference in hydraulic head between these two points. Water will move from the point represented by the high level towards the low. The difference in head, ΔH, divided by the distance between the points, Δl, is the hydraulic gradient. It is known empirically from the work of Darcy in 1856, that the volume, v, of water flowing across unit area of cross-section of soil in unit time is proportional to the hydraulic gradient, i.e.,

$$v = k\,\Delta H/\Delta l \qquad \text{- - - - - - - - - (9)}$$

The constant, k, from analogy with similar equations for flow of heat and electricity, has been termed the hydraulic conductivity (Richards, 1952). It is commonly expressed in units of cm/sec or in/hr. For use with the one fluid (water) within the usual range of temperature found in soils, this equation is satisfactory. However, the value of k will vary not only with the properties of the soil but also with those of the fluid. It is desirable therefore to have also a permeability constant which will reflect only the properties of the soil. In this case

$$v = -k'\eta^{-1}d\Phi/dl \qquad \text{- - - - - - - - - (10)}$$

where η is the viscosity of the fluid, Φ is the potential per unit volume of fluid in ergs/cm^3, and k' represents the intrinsic permeability of the soil in cm^2. The terms hydraulic conductivity (k) and intrinsic permeability (k') are those proposed by the Sub-committee of the Soil Science Society of America on Permeability and Infiltration (Richards, Chairman, 1952). In addition, this sub-committee proposed that the word permeability should continue to be used qualitatively to represent the quality or state of a porous medium relating to the readiness with which it conducts or transmits fluids.

During flow there is a loss of energy in overcoming frictional resistance within the pores. The rate of change of energy in the direction of flow is given by $-d\Phi/dl$ in which the potential, Φ, is the sum of the pressure or capillary component, ψ, and the gravitational component, φ (see Section 2 vi). When potential due to pressure is expressed in cm of water and that due to position is in cm above a datum, this total becomes the head. The gradient in total potential or head represents the driving force and Darcy's law simply states that flow is proportional to this. Darcy's law is applied to the flow of water through unsaturated as well as saturated soil using tensiometers instead of open tubes for measuring potential. In this case the conducting system for liquid water becomes progressively less effective as the soil becomes drier and permeability for a given soil is therefore found to vary with water content.

There is a great range in the permeability of soils. For example the porous materials shown in Fig. 11 have intrinsic permeability ranging from about 10^{-11} to 10^{-6} cm^2 and a comparable range is met with in soils. Because of this and because of the variety of units used by various workers, some reference values have been given in Table 5 as an aid in interpreting published values of permeability. In this table the descriptive terms and the corresponding permeability ranges are those used by O'Neal (1952) for saturated subsoils.

Smith and Browning (1946) have given a range of standards in which the classes differ by a factor of 10 and they have provided comments on the associated water holding and aeration conditions. These are summarized in Table 6. The effects of cracks are not allowed for in either Table 5 or 6.

TABLE 5

Permeability classes of O'Neal (1952) *for saturated subsoils and the corresponding range of hydraulic conductivity and of intrinsic permeability*

Class	*Hydraulic conductivity, k* 10^{-5} *cm/sec*	*in/hr*	*Intrinsic permeability, k'* 10^{-10} cm^2
Very slow	<3	<0.05	<3
Slow	3–15	0.05–0.2	3–15
Moderately Slow	15–60	0.2–0.8	15–60
Moderate	60–170	0.8–2.5	60–170
Moderately rapid	170–350	2.5–5.0	170–350
Rapid	350–700	5.0–10.0	350–700
Very rapid	>700	>10.0	>700

TABLE 6

Permeability classes of Smith and Browning (1946) *for saturated subsoils*

Class	*Hydraulic conductivity, k, in/hr*	*Comments*
Extremely slow	<0.001	So nearly impervious that leaching process is insignificant.
Very slow	0.001–0.01	Poor drainage results in staining; too slow for artificial drainage.
Slow	0.01–0.1	Too slow for favourable air-water relations and for deep root development.
Moderate	0.1–1	Adequate permeability.
Rapid	1–10	Excellent water holding relations as well as excellent permeability.
Very rapid	>10	Associated with poor water holding conditions.

II. Permeability of Saturated Soil

A great deal of work has been done on the permeability of saturated soil. There have been two quite different objectives in all this—(i) to find out how water movement is affected by various properties of the soil and of the solution moving through soil and (ii) to determine for a particular site how freely water will move through the soil in connection with irrigation, drainage, or erosion problems. From the practical standpoint, measurements on saturated soil are appropriate chiefly to drainage or other ground water problems but they also give indirect information about structure and structure stability (and hence about water entry) in soils which may not commonly be saturated under field conditions.

In principle, the measurement of permeability in the laboratory is simple; but in fact there are many difficulties in applying the results quantitatively in the field. Water is passed through soil held usually in a cylindrical container which either fits closely to the soil or is sealed to it by paraffin wax to prevent passage of water down the side of the soil column. The difference in head between two levels and the rate of flow are measured and the permeability can then be calculated from Darcy's law either in the units of hydraulic conductivity or intrinsic permeability. Either "undisturbed" cores or artificially packed soil can be used.

Cores may be obtained in the field using special augers with hand or machine operation or by pressing a core cutter into the soil. Sampling is usually done vertically but it may also be done horizontally in horizons which are not isotropic with respect to permeability. Cores may then be coated with paraffin wax and sealed with the wax to a metal cylinder (e.g., Slater and Byers, 1931) or to a metal sheet wrapped around the core at each end (Bloodworth and Cowley, 1951). Alternatively the permeability may be measured with the sample still retained in the cylinder which formed the sampling tool or in a liner to that tool (e.g., Bower and Peterson, 1950). In this case no paraffin wax is used. Methods for taking cores and using them for permeability measurements are reviewed by Reeve (1957a).

Undisturbed cores represent field conditions better than broken and repacked soil but they require much care in sampling, handling, and assembling to avoid cracking of the soil or channeling between the core and its container. Because of this difficulty, ground up soil is sometimes used and is packed by a reproducible procedure into a cylinder. The permeability of this material may differ greatly from that of the natural soil the structure of which may have been grossly altered, but nevertheless the relative effects of some soil properties such as degree of sodium saturation will appear in these measurements. Fireman (1944) has given details of such a procedure and has examined all aspects of it critically.

Bodman (1938) showed that permeability measurements on packed columns in the laboratory may be much affected by the amount of water that has passed through in making the test. Three stages were apparent: permeability decreased at first, then it increased, and finally it decreased slowly. Bodman attributed the overall decline to the progressive removal of soluble salts by the water and the consequent dispersion and rearrangement of particles. Christiansen (1944) showed that, at the same time as this process is going on, air which has been entrapped in the pores during wetting is gradually removed in solution. This frees those pores which have been blocked with air and so increases the permeability during stage 2. Greacen and Huon (1953) have presented sets of microphotographs of beds of aggregates through which water is passing. These show a progressive increase in the volume of the aggregates (and an accompanying reduction in size of the large pores) and also during stage 2 the gradual removal of air bubbles. Actual movement of dispersed particles with the stream was not observed.

According to Allison (1947) microbiological activity may be responsible for the progressive decrease in permeability during measurement, but this was not considered to be the cause in the experiments of Bodman and Fireman (1950) and of Greacen and Huon. In the case of the beds of aggregates of Greacen and Huon the entire runs occupied only a few hours, too short a time for microbiological activity to cause blocking of the pores.

The question of what solutes to use in water which is being passed through soil during a permeability measurement is a difficult one. For irrigation investigations, Fireman recommends using the water which will be applied in practice to the land. In these circumstances the hydraulic conductivity of a sample should not normally be lower than 0.1 cm/hr (3×10^{-4} cm/sec) if the soil is to be successfully used in irrigation culture according to the experience of the U.S. Salinity Laboratory (Richards, 1954) based on measurements made on disturbed samples.

The concentration and composition of salts dissolved in irrigation water can affect the permeability of a particular soil greatly by affecting the swelling and dispersion of the clay. If the water contains much sodium ion, the proportion of exchangeable sodium to other exchangeable cations may increase as the solution passes through and permeability will decrease because clay with much exchangeable sodium disperses readily. On the other hand the decrease may be prevented if the salt concentration is sufficient to prevent dispersion and, as shown by Fireman and Bodman (1940) and Bodman and Fireman (1950), permeability may then depend more on salt concentration in the water than on the exchangeable cations on the soil. Hence permeability can be maintained irrespective of the degree of sodium saturation of the exchange complex provided the concentration of salts is sufficiently high. For this reason it is often advisable in reclaiming a saline soil containing much exchangeable sodium to apply gypsum (or other amendment) before too much of the naturally occurring salt has been removed by leaching. This is discussed in relation to irrigation by Richards (1954) and, in relation to soils flooded by sea water,

by Deloffre (1947, 1953) and Beekom *et al* (1953). Vulnerable soils in which exchangeable sodium comprises about 15 per cent or more of the cation exchange capacity are referred to as sodic soils.

For any given percentage of sodium in the exchange complex, Quirk and Schofield (1955) sought a corresponding salt concentration which was high enough for permeability to be maintained (Fig. 8). This means that in reclaiming a saline sodic soil, the salt concentration of the irrigation water should be kept above the level appropriate to the particular content of exchangeable sodium in the soil. Quirk and Schofield considered that the relation given in Fig. 8 should apply to most semi-arid soils but recognized from the work of Emerson (1954) and others that it cannot be applied universally. Indeed Emerson suggests that the stability of aggregates can be measured by finding the concentration of NaCl solution at which the permeability of a bed of aggregates which have been Na-saturated falls almost to zero when a solution of decreasing concentration has been passed through. The permeability of beds made of stable aggregates from grassland did not reach this stage until the concentration was greatly lowered (0.3×10^{-3}N). In contrast to this, less stable aggregates from a permanent mangold field nearby broke down and the permeability became effectively zero when the concentration was 34×10^{-3}N.

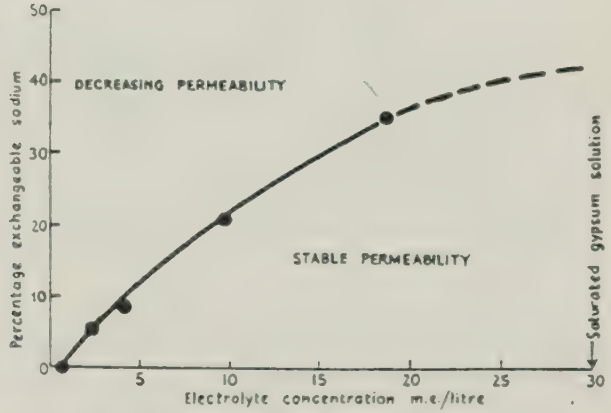

FIG. 8. *Concentration of electrolyte required at different degrees of sodium saturation if permeability is to be kept stable (Quirk and Schofield, 1955)*

III. Field Measurements of Permeability of Saturated Soil

Permeability may be determined in the field on soil horizons lying below a water table using methods developed by Hooghoudt, Ernst, Kirkham, Childs and their co-workers. The methods depend upon the alteration of the level of the water surface in bore holes by removal or addition of water. Van Bavel and Kirkham (1948) have used unlined auger holes about 4 inches in diameter for determining permeability in this way. Water is removed from the hole and the rate at which the water level then moves upward towards its original position is measured. The "auger hole" method measures horizontal permeability primarily and all saturated horizons penetrated by the hole are involved. Hence the result may be dominated by one permeable horizon in non-uniform stratified material. The method is therefore of most value in unstratified soils. The hydraulic conductivity, k, may be determined for the "auger hole" method by

$$k = [\pi r^2/(d - z)\,A] \quad (\Delta z/\Delta t) \qquad \text{- - - - - - - - - - (11)}$$

in which r is the radius of the hole, z is depth of water in the hole, d is the depth of the hole below the water table, t is the time for water to rise a distance Δz in the hole, and A is a geometrical constant with dimensions of length.

If a closely fitting liner is forced in while a hole is being bored and only a restricted section near the bottom is left as an unlined cavity free to admit water, then the permeability of any particular horizon lying beneath the water table can be measured more specifically. Luthin and Kirkham (1949) have developed a method, which they refer to as the "piezometer method", for doing this. They used a narrow one-inch diameter cavity about 4 inches long and found that in this case they were measuring horizontal permeability principally (see Fig. 9). The wider the hole and the shorter the length of the cavity left unlined, the more nearly the measurement becomes one of vertical permeability and in the "tube method" of Frevert and Kirkham (1948), the extreme case of this is reached with the liner going right to the bottom of a wide (8 inches diameter) hole which the water can enter only through the base. In this case vertical permeability is principally measured (Fig. 9). The shape of the holes and the "sampling" volume (as represented by the zone in which 80 per cent of the hydraulic head is dissipated) are sketched in Fig. 9 which is adapted from diagrams of Reeve and Kirkham (1951).

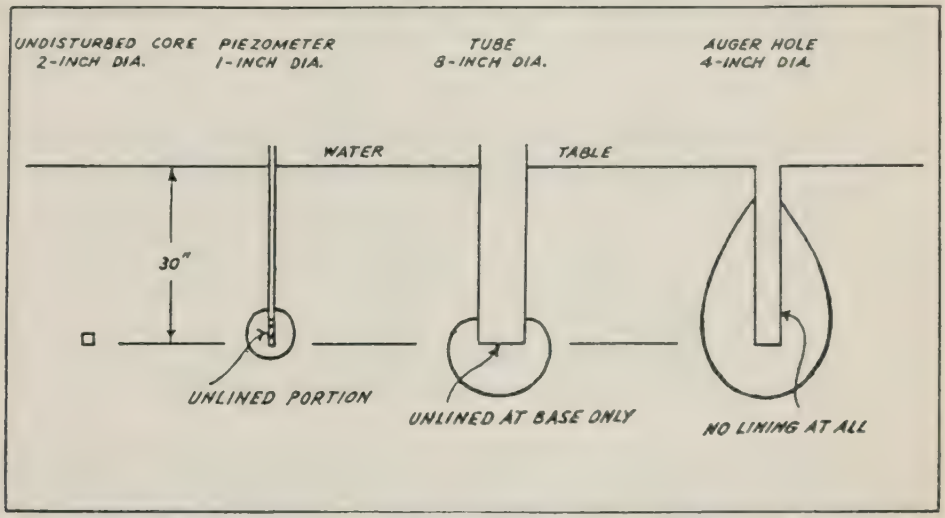

FIG. 9. *Comparison of four permeability methods showing the relative sizes of the "samples" represented in each. In the field methods the sample zones are for the volumes of soil through which 80 per cent of the hydraulic head is dissipated (Reeve and Kirkham, 1951)*

For piezometers and tubes, hydraulic conductivity is given by

$$k = \{\pi r^2/[A(t_2 - t_1)]\} \ln (y_1/y_2) \qquad \text{- - - - - - - - - - (12)}$$

in which y_1 and y_2 are the distances from the water level in the hole to the water table at times t_1 and t_2 respectively, r is the radius of the hole, and A is a geometrical constant with dimensions of length.

Details of how to sink holes with liners, how to overcome (by preliminary pumping) the puddling of the exposed surface caused by the boring operation, and the use of Equations (11) and (12) or modifications of them are discussed by Reeve and Kirkham (1951) and Luthin (1957b).

Childs (1952) has developed a method for measuring permeability which depends upon the pumping of water from one hole to another. Measurements are made of the rate of flow through the pump and the distance by which the original water level is lowered in one hole and raised in another. Childs, Cole and Edwards (1953) used holes of 6 inches and 8 inches in diameter and 3 feet apart in the field and lined them with gravel held in place by wire netting to prevent collapse of the soil. The hydraulic conductivity is given by

$$k = \frac{\chi \, dQ/dt}{(l + l_f)\Delta H}$$

where dQ/dt is the rate of flow through the pump when the difference between the water levels

of the two wells is established as ΔH, l is the depth of the wells below the water table, l_f is an end correction factor (20 cm for isotropic soil), and χ a geometrical factor (0.735 for wells 1 m apart and 20 cm in diameter). According to Kirkham (1955) the method has two possible defects —it requires more man-hours because of the placing of the screen and there is the chance of clogging when water has to go inwards.

The two-well method measures horizontal conductivity k_h. Childs *et al* show that if in addition a measurement is made of the apparent conductivity, k_a, by the method of Luthin and Kirkham, then the vertical component, k_v, can be calculated from

$$k_a = (k_h k_v)^{\frac{1}{2}}.$$

Childs, Collis-George and Holmes (1957) found that horizontal permeability differed greatly from that in the vertical direction at certain sites. An example from their data is given in Table 7. Most of their soils were of alluvial origin so that anisotropy could be expected. However, at more than half their sites the soils were isotropic and Childs (1957) considered that appreciable anisotropy may not be as common as is sometimes believed.

TABLE 7

Hydraulic conductivity of some subsoil horizons at Romney Marsh, England

(From data of Childs, Collis-George and Holmes, 1957)

Site	Texture	Hydraulic conductivity, cm/sec	
		Horizontal	Horizontal/Vertical
1	Sandy clay	5.8×10^{-4}	1
2	Sandy loam to fine sand	1.6×10^{-3}	1
4	Clay	1.1×10^{-4}	0.03

The advantages of field methods for determining permeability are that the soil is less disturbed and that the volume of soil examined is much larger and hence is less liable to sampling error than the laboratory samples. If the permeability data are to be used for drainage design, there is also less difficulty over air in the pores and solutes in the water since the soil is examined under moisture conditions which have been established naturally. In some cases permeability may later change after artificial drainage has been introduced because of changes in these and other conditions, but the initial basis for design should be sound.

The main disadvantage of these field methods is that they are limited to sites and horizons where the soil lies below a water table. This restricts their use and they are therefore of main value in relation to drainage design in areas already troubled with waterlogging. In this they have much promise. In other circumstances the laboratory procedures can be used. Although they are not as well suited for representing permeability quantitatively for direct field use (as in drainage design), they have a place in work done on the effects which imposed treatments and altered properties have upon permeability.

IV. Permeability of Unsaturated Soil

Saturation is an undesirable condition for all except specialised crops such as rice. For adequate aeration, there should be an appreciable amount of air-filled pore space and, if it occurs at all, saturation does not persist for long in most soil horizons in which plant roots are active. Hence gains and losses of water very commonly involve movement through soil which is unsaturated to some degree. Such movement can occur in both the liquid and vapour phases and the relative importance of the two mechanisms depends amongst other things upon the degree of saturation of the soil.

Permeability for water (as a liquid) decreases as soil becomes drier because the introduced air acts as a non-conducting part of the system. This was first demonstrated quantitatively by Richards (1931) using a method by which water could be maintained at a constant mean suction (and water content) while a known gradient in total potential was applied to it. The permeability corresponding to the chosen mean suction was then calculated from the steady rate of flow of water through the soil by means of Darcy's law. In the original method of Richards, known suctions were applied at the ends of the small test column through suction plates and the potential gradient was measured by means of tensiometers. Later, gas pressure instead of suction was used by Richards and D. C. Moore (1952) to control the unsaturation of the column. Using a different method, R. E. Moore (1939) measured the potential gradient in large soil columns through which water was passing and evaporating at one end. Childs and Collis-George (1950) measured the permeability of unsaturated materials packed in a column 3 meters long and of adjustable inclination. This column was long enough to provide a zone of uniform water content and suction in which potential gradient was known from the inclination of the column.

In all these methods the attempt is made to reach steady state conditions in which the inflow rate is equal to the outflow rate. It is also possible to measure unsaturated permeability while soil is in the process of wetting or drying provided data for potential gradient and flow velocity can be obtained with sufficient reliability under these conditions. S. J. Richards and Weeks (1953) have adopted this approach using a column of wet soil from which water was continuously withdrawn under suction. Similarly L. A. Richards, Gardner and Ogata (1956) measured the suction and water content at various depths and times while water was evaporating from a deeply irrigated soil. In this way they were able to determine unsaturated permeability in the field. Ogata and Richards (1957) were able to reproduce these results when the same soil was covered against evaporation and water moved downward after irrigation instead of upward. The two sets of results were also found to agree rather well with those obtained on the same soil by Richards and Moore (1952) in the laboratory under steady state conditions. However this particular comparison is not as critical as the other because a packed column of fragmented soil was used in the laboratory and this may have been of different structure from soil in the field.

In these non-steady state methods, suction was measured by means of tensiometers. This limits the degree of unsaturation which can be handled because these instruments cannot be used at suctions greater than about 850 cm. Staple and Lehane (1954) worked on the permeability of soil drier than this by following the movement of water from moist to dry soil. They inferred the suction from the measured water content at different times and depths in their columns using a drying curve for the relation between water content and suction. Since wetting conditions were also represented in their experiment, their results are not free from the effects of hysteresis. Also vapour transfer to the dry soil could be appreciable in their short columns. These effects are recognized by the authors and they probably limit the accuracy of their permeability data especially at the dry end of their range of values. Gardner (1956) has also measured permeability in relatively dry soil. He did this in the pressure plate and pressure membrane apparatus from data obtained on the progressive loss of water from the soil under each of a number of small increments of gas pressure. In this method it is necessary that imperfect contact between soil and the supporting membrane should not impose an important resistance to flow.

In all these measurements, Darcy's law has been used to obtain the permeability. For Darcy's law to be valid in these unsaturated conditions, the constant k or k' in Equations (9) and (10) should be unaffected by the magnitude of the potential gradient used in making the measurements. Evidence that this is the case (as it is in saturated materials) is provided in the data of Childs and Collis-George (1950) one set of which is shown in Fig. 10. Here the permeability is clearly unaffected by the gradient used when making the measurements. Fig. 10 also gives some indication of how permeability decreases with decreasing water content. This particular curve applies to a sand whose particles are between 0.5 and 1 mm in diameter. The size distribution of the pores in this material is given in Fig. 11 as curve 1a.

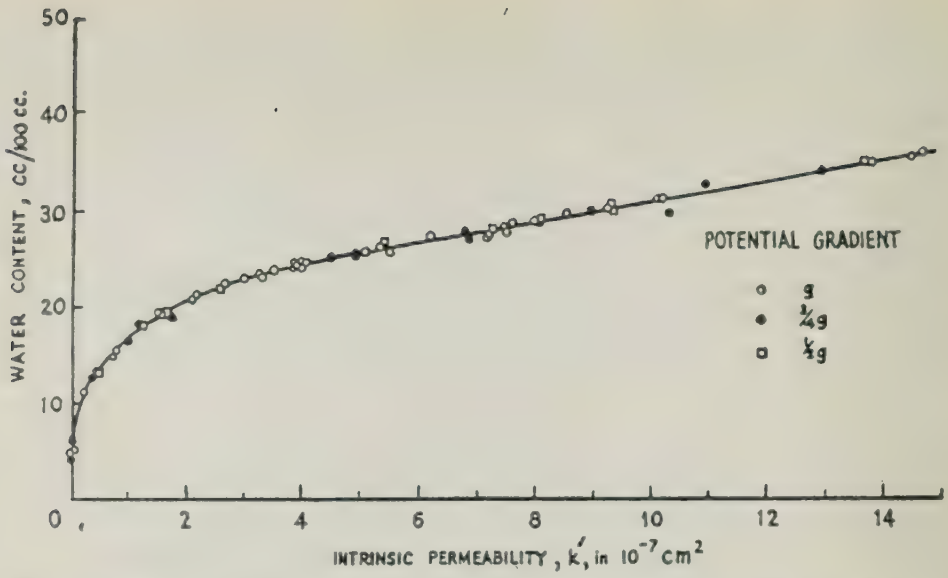

FIG. 10. *Permeability of unsaturated sand measured at hydraulic gradients of* 1, ¼ *and* ½ (*Childs and Collis-George*, 1950)

The permeability at any given water content or suction can be very different in different materials. Because of this and because permeability is difficult to measure in unsaturated soil, much attention has been given to ways of calculating it from other data. Some general principles governing the relation of permeability of saturated and unsaturated materials to these other properties will next be discussed.

5. PERMEABILITY AND PORE SPACE

I. Introduction

Easily recognizable features of texture and structure are commonly used as clues for assessing the permeability of soils qualitatively in the field (Bendixon, Hershberger and Slater, 1948; O'Neal, 1952). The depth and spacing of drains for example is usually worked out by examining the texture of the profile in inspection bores, and Reeve (1957a) considers this is often sufficient when coupled with drainage experience in the area concerned.

Talsma and Flint (1958) found that texture had an important influence on permeability in deep subsoils but they also noted that permeability decreased with depth due to the compressing effect of overlying soil. The effect of texture is most pronounced in coarse materials of fairly uniform particle size because here systems with large particles have also large pores to conduct the fluid. In this connection, Lutz and Leamer (1939) and Pillsbury (1950) found that permeability increased exponentially with size of particles in uniformly sized sands. However in soils of finer texture, this trend does not always hold and Bodman (1936) found that texture of soils finer than fine sandy loam was unrelated to permeability in a group of soils whose bulk density was greater than 1.4 g/cm^3. Similarly it is a matter of common observation that some soils can be highly permeable even if their clay content is high. In this case the arrangement of the particles and the consequent sizes of the conducting pores cannot be inferred from the texture and any field assessment of permeability has to be made from observable features of the structure.

It is recognized that, in general, permeability will increase with porosity and with the size of the pores through which the fluid has to pass and permeability equations have been developed on that basis. Pore size has been measured in indirect ways to provide the data required for these equations. One way has been to infer the geometry of the pore space from the size and packing of the particles which form the framework of the pores. Another has been to calculate the size distribution of the pores from a curve relating water content to suction.

II. The Effect of Particle Size and Porosity on the Size of the Conducting Channels

The earliest attempts were made in the first of these ways, at a time when measurement of the size distribution of fine pores had not yet been undertaken by suction methods. Classical work on the geometry of the pore system within beds of spherical particles of uniform size was done by Slichter (1899). He found that flow occurred through sinuous pore channels which were alternately of large and small section but that by making compensating approximations it could be assumed to occur in straight capillaries of a uniform circular section whose area was that of the smallest parts of the channel. He was then able to derive an equation which related permeability to porosity and particle diameter in an "ideal" porous system. There were flaws in Slichter's analysis and Green and Ampt (1912) subsequently found experimentally that his equation could be applied very well to beds of sand grains of irregular shape but that it was less accurate when applied to the ideal system for which it was derived. Smith (1932) noted that Slichter had erred in assuming as he did in his analysis that the ratio of an average to a minimum section of the pore channel was independent of porosity. Smith developed an equation based on the mean triangular section which fitted Green and Ampt's data satisfactorily but at the same time he found that the data for irregularly shaped sand grains were fitted still better if he used the minimum rather than the mean triangular section. Graton and Fraser (1935) discussed the possibility that a relation might be found between permeability and size of the straight duct common to all pores in a given line of flow through spheres packed in a regular fashion.

The next step to be taken was the use by Kozeny (1927), Fair and Hatch (1933), and Carman (1937) of surface area of particles and porosity as the basis for a permeability equation. The

derivation and application of the Kozeny equation is discussed fully by Carman (1939, 1956). It is developed from Poiseuille's equation for the viscous flow of fluids through narrow tubes which may be written as

$$u' = (r^2 \Delta\Phi/\Delta l)/8\eta \qquad \text{- - - - - - - - - (13)}$$

where u' is the mean velocity of flow in cm/sec through a tube of radius r and length Δl, η is the viscosity of the fluid in poises and $\Delta\Phi$ is the difference in potential between the ends of the tube in ergs/cm^3.

To apply this equation to flow through a porous system, allowance is made for the fact that the fluid does not follow a path of length Δl but a more tortuous one of greater length Δl_e. Also since the pore space through which the fluid passes, represents only a fraction, ε, of the total volume of porous material, the actual velocity, u', in the tube is given by (u/ε) $(\Delta l_e/\Delta l)$ where u is the apparent velocity through the porous material. The potential gradient becomes $\Delta\Phi/\Delta l_e$ along the longer path. The radius, r, of a tube of circular section in Poiseuille's equation is replaced by $2\varepsilon/S$ where S is the surface area per unit volume of the bed, and ε/S is equivalent to the volume of the tube divided by the area of its walls. But since the "tubes" do not have a circular cross-section, a shape factor, a_o, is introduced as an empirical constant (using the device of the "hydraulic radius") to take the place of a factor 2 which would otherwise appear in the equation for tubes of circular section. With these substitutions in Poiseuille's equation, the apparent velocity for the porous medium as a whole becomes

$$u = \left(\frac{\varepsilon^3 \Delta \Phi}{S^2 a_o \eta \Delta l}\right) \left(\frac{\Delta l}{\Delta l_e}\right)^2$$

Also since $u = k'\eta^{-1}\Delta\Phi/\Delta l$ from Darcy's law, k', the intrinsic permeability, is given by

$$k' = \varepsilon^3/S^2 a \qquad \text{- - - - - - - - - (14)}$$

where a is a constant usually referred to as the Kozeny-Carman constant.

The treatment given here follows that of Carman (1937) who interpreted a as being the product of a shape factor, a_o, and a tortuosity factor $(\Delta l_e/\Delta l)^2$. Carman has assigned a value of 5 to a from experimental evidence and has taken a_o as equal to 2.5 and the tortuosity $\Delta l_e/\Delta l$ as equal to $\sqrt{2}$ for most purposes. This value for $\Delta l_e/\Delta l$ will be correct if, due to tortuosity, the fluid can be considered to move at an angle of 45° to the direction in which flow is being measured.

The Kozeny equation has been widely used in fields concerned with the flow of water, petroleum and gases through porous materials. The extensive literature which has been developed around it has been fully reviewed by Carman (1956). The equation has been particularly useful in material with pores of fairly uniform size but it has been shown to fail seriously under other conditions. In soil, for example, where fairly large pores can occur although the material is made up of fine particles, the surface area of the individual particles making up the aggregates need bear little relation to the permeability. There are other objections to the equation even when applied to uniform materials because tortuosity and the Kozeny-Carman constant, a, have been shown by the experiments of Wyllie and Rose (1950) and Wyllie and Gregory (1955) to vary inversely with porosity, ε. Marshall (1958) has also shown that, if the Kozeny equation is to be reconciled with Equation (17), a has to take the value of $2/\varepsilon$. Wyllie and Gardner (1958b) consider the probability of pore interconnections to be more important than the sinuous path in restraining flow and they have therefore treated tortuosity as a function of porosity. In an equation which they have derived they have retained the framework of the Kozeny equation but the connection is only a slight one and they propose that the term impedance might be used in place of tortuosity.

III. The Effect of Size Distribution of Pores on Permeability

Because of the limitation of equations based on particle size or surface area, many workers have turned to pore size distribution as the basis for developing a permeability equation. In this work pore size has been measured indirectly by suction methods chiefly. Baver (1938) recognized that permeability should vary with the fraction of the apparent volume occupied by large pores and inversely with suction required to empty these pores. He sought a basis for selecting a critical suction to define the "large" pores from the shape of the curve relating water content to the

logarithm of the suction. Nelson and Baver (1940) tried suctions of from 10 to 180 cm and found that the water withdrawn at 40 cm suction gave the best relation with permeability. Smith, Browning and Pohlson (1944) obtained their best relation when they considered together three classes of pore size—those emptied at 10 cm suction, those emptied between 10 and 40 cm, and those between 40 and 100 cm. The volume of the pore space in each size class was weighted in favour of the large pores—the volume in the second class being divided by 4 and that in the third by 10. This basis for weighting was arrived at empirically without recourse to Poiseuille's law. They found that the third class made relatively little contribution except in soils of low permeability.

Purcell (1949) working with porous rocks used an empirical equation relating permeability to the size distribution of the pores. The pore system was considered as being made up of straight parallel capillary tubes, each of circular section and each making a separate and independent contribution to flow according to its radius. Purcell's equation which is developed with the aid of Poiseuille's and Darcy's laws may be expressed as

$$k' = 7 \times 10^{-3} F \, \varepsilon \int_{c=0}^{c=\varepsilon} h^{-2} dc$$

where k' is the permeability in cm^2, ε is the porosity in cm^3/cm^3, c is the water content in cm^3/cm^3 corresponding to a suction of h in cm of water and F is an empirical factor. Purcell tested his equation on materials for which both pore size distribution and permeability had been measured and found that his factor, F, had to vary from 0.07 for samples of low permeability to 0.36 for samples of higher permeability if calculated and measured permeability were to agree. He anticipated that other materials might require other values for F and this is likely since the range of permeability covered was not very large. The size distribution and permeability of some of the rocks used by Purcell are illustrated in Fig. 11 and Table 8. Pressure membrane data are represented here in preference to mercury injection data which Purcell also used for determining size distribution. Wyllie and Spangler (1952) used an equation similar to Purcell's except that the Kozeny-Carman tortuosity factor appearing in it in place of Purcell's factor was determined by the electrical conductivity method. This method gave values for the factor which were very different from the basic Kozeny-Carman constant but permeability calculated by this means agreed well with measured permeability. Burdine (1953) has introduced a further modification to this form of equation by allowing empirically for variation of the tortuosity factor with porosity and Wyllie and Gardner (1958a; 1958b) have provided a derivation for an expression similar to Burdine's. As noted above they have introduced the term impedance factor in place of tortuosity factor in their equation.

Childs and Collis-George (1950) have presented an equation which is based on the probability of occurrence of all possible pore-size sequences. At any cross-section where a porous medium is considered to have been broken and rejoined randomly, the fraction of the area of the section devoted to a sequence of pores of one particular size followed by those of some other size can be calculated from the size distribution of the pores. The assumption is made that in such a sequence, the resistance to flow is confined to the smaller of the two pores. From Poiseuille's law, flow through this pore is proportional to the 4th power of its radius and, since for a given size distribution the number of such pores in the cross-section is inversely proportional to their radius, flow across this fraction of the section is proportional to the square of their radius. The permeability, k', due to all possible sequences is then given by

$$k' = M \sum_{\rho=0}^{\rho=R} \sum_{\sigma=0}^{\sigma=R} \sigma^2 f(\rho) \delta r f(\sigma) \, \delta r \qquad \text{- - - - - - - - - (15)}$$

in which σ and ρ are radii of two pores forming a sequence, σ being the smaller, $f(\sigma)\delta r$ and $f(\rho)\delta r$ are the fractions of the cross-sectional area taken up by pores of each of these sizes, R is the radius of the largest pore concerned with flow and M is an empirical constant.

A value for the proportionality factor M in this equation was obtained experimentally by matching with the measured permeability of a water-saturated sand. The equation was tested on this and two other materials against measured permeability at various degrees of saturation in which cases the size distribution of only those pores which are saturated forms the basis of calculation. The comparison was satisfactory but each of the materials (Fig. 11, curves 1a, 1b,

and 4) had a very narrow range of pore size, and the test made on the matching factor was therefore a limited one. Nevertheless the equation represented a considerable step forward because not only was the whole of the size distribution of the pores represented but also the effect on permeability of various sequences of pores of different sizes was allowed for.

The fit of one pore to the next can be looked at in another way. When one pore passes into the next in line in the apparent direction of flow, the alignment is not necessarily perfect. The projection of one on the other gives the area common to each which will be considered to represent the size of a connecting neck. An average size of these necks can then be obtained (in a form suitable for substitution in Poiseuille's equation) from the size distribution of the pores. It is given (Marshall, 1957; 1958) by

$$r_t^2 = \varepsilon_n^{-2} [r_1^2 + 3r_2^2 + 5r_3^2 + \ldots\ldots + (2_n - 1)r_n^2] \text{ - - - - - - (16)}$$

where r_t is the radius of a tube of circular cross-section having the same cross-sectional area as this average neck, ε is the porosity in cm^3/cm^3, and r_1, r_2, r_3, ... and r_n cm represent the mean radius of the pores (in decreasing order of size) in each of n equal fractions of the pore space. Flow across any plane normal to the apparent direction of flow is considered to occur through tubes of radius r_t on the assumption that flow is controlled by neck size rather than pore size. The whole of the pore space is considered to contribute, thus allowing for the multiple outlets available to the larger pores. With the aid of Poiseuille's law (Equation 13) and Darcy's law (Equation 9) the following permeability equation can then be derived.

$$k' = \varepsilon^2 n^{-2} [r_1^2 + 3r_2^2 + 5r_3^2 + \ldots . + (2n - 1)\, r_n^2] / 8 \text{ - - - - - - (17)}$$

where k' is the intrinsic permeability in cm^2 and the other symbols are as given for Equation (16). Equation (17) is related mathematically to Equation (15) although they arise from different models. It can be adapted for use in unsaturated materials. If a material contains c cc water per cc of apparent volume, then c is substituted for ε and n refers to the water-filled pore space.

Using Equation (17), permeability can be calculated from the size distribution of the pores. Since most pore size measurements are made indirectly by suction methods, it is convenient to use the suction instead of the pore radius r in these calculations. From Equation (3), in Section 2(III), r^2 is given by $2.25 \times 10^{-2} h^{-2}$ when suction, h, in cm of water is measured, using water at 20°C. Substituting in Equation (17),

$$k' = 2.8 \times 10^{-3}\ \varepsilon^2 n^{-2} [h_1^{-2} + 3h_2^{-2} + 5h_3^{-2} + \ldots . + 2(n-1)h_n^{-2}] \text{ - - - (18)}$$

where h_1, h_2, ... and h_n represent the suction in the equal classes; h_1 (corresponding to r_1) belongs to the class with the largest pores and h_n (corresponding to r_n) belongs to that with the smallest. Calculations are easily made from a water content-suction curve and detailed examples are given by Marshall (1958).

Permeability is obtained as intrinsic permeability in units of cm^2 in Equations (17) and (18). When dealing with flow of water, it can instead be obtained in the units cm/sec as hydraulic conductivity, k, if the numerical constant in Equation (18) is multiplied by g/η and thus becomes 2.7×10^2 at 20°C.

In the derivation of Equation (17), it is assumed that, in porous materials of small pore size in which restrictions in the flow path occur at short intervals, flow is dependent on an average size of the connecting necks. This assumption of a controlling neck size will result in an underestimate of permeability but this is offset to a degree by assuming necks to be circular in section, by neglecting any deviation of the actual direction of flow from the apparent direction, and (in the calculation of an average neck size) by allowing for only one neck at the end of the smaller pore in each sequence. There are arbitrary features in the calculation of neck size and in the assumption of a uniform potential gradient in the pores (See Section 8 1).

IV. Comparison of Calculated and Measured Permeability

The generality of an equation based on these assumptions has to be examined by testing against experimental data. This has been done using the materials whose pore size distribution is illustrated in Fig. 11. Measured and calculated permeabilities for these materials are compared in

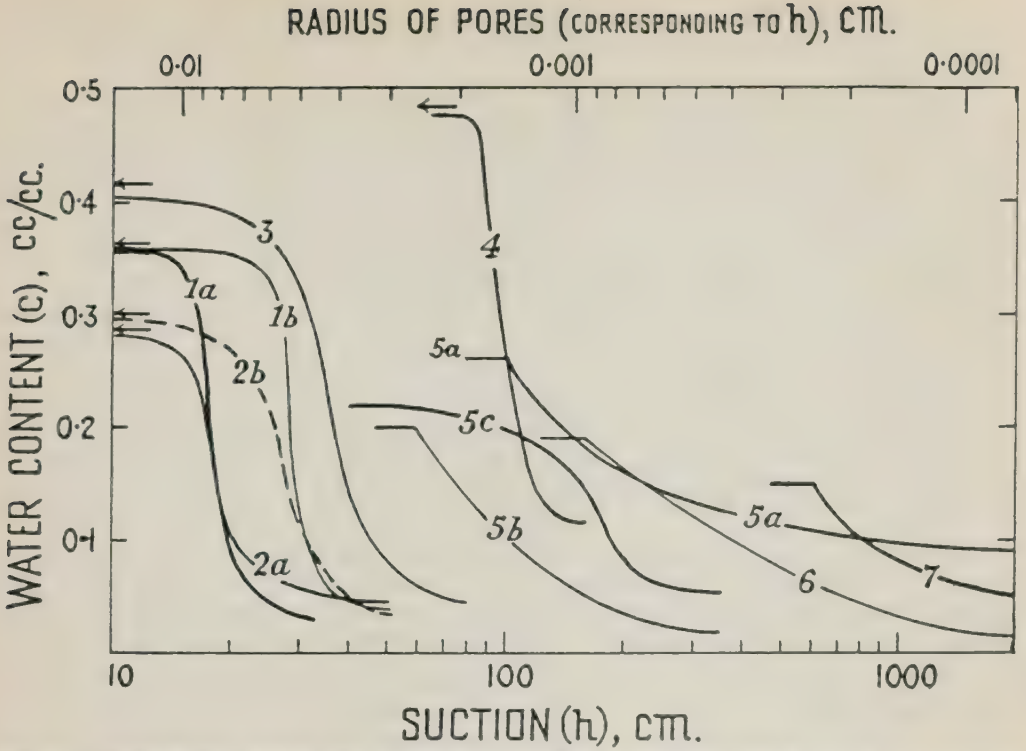

FIG. 11. *Relation between water content, suction, and pore size in various non-swelling porous materials (drying curves). The sources of data are given in Table 8. The water content corresponding to the upper horizontal portion of the curves or to the arrows represents pore saturation for each material*

Table 8. The agreement is encouraging considering the wide range of materials represented. These include porous rocks of low porosity (No. 5, 6, 7) as well as uncemented sands of high porosity (No. 1, 2, 3, 4) and materials with widely distributed sizes of pores (No. 5a, 7) as well as those with narrow distributions (No. 1, 4). The results come from measurements in unsaturated (No. 1, 3, 4) as well as water-saturated materials (No. 1, 2, 3, 4) and from materials in which air was the fluid used for measuring permeability (No. 5, 6, 7). The only exceptional case in the original comparison (Fig. 2 of Marshall, 1958) was for Oakley sand which has similar properties to material 3 of Fig. 11. It will be found on referring to the data of Moore (1939) for Oakley sand that permeability was constant between saturated and about half-saturated conditions. This constancy of permeability over such a range of water content is most improbable and also the magnitude of the permeability near saturation is much lower than that of comparable sands. Hence the data for Oakley sand have not been included in Fig. 11 and Table 8.

Table 8 serves to indicate that pore size data, obtained indirectly from the curve relating water content to suction for a porous material, can provide a real basis for calculating permeability of sands and porous rocks. Individually the results are not highly accurate, but a great range of permeability is covered and there is much diversity of pore space and pore size in the materials. However the approximations and assumptions which have to be made will not be valid in all circumstances. In the first place, error will occur when the pore size distribution is not reliably determined by the suction data as is the case for example in swelling soils. Another source of error will be present when the pores are not randomly distributed. If the particles are

preferentially oriented or sorted, permeability in one direction may exceed that in another because the pores are so arranged as to favour better pore to pore connections in one direction than in another.

The evidence that pore size is consistently related to permeability over a wide range of size gives added point to the methods of water displacement (under both suction and gas pressure) that were used in measuring the size distribution of the pores. However none of the data of Fig. 11 were obtained by the method of vapour adsorption. It is of interest therefore to turn to a size distribution curve obtained by means of the Kelvin equation by Carman (1951) from the adsorption of freon on a compressed plug of Linde silica with pores smaller than 100 Å in radius. From this curve the permeability of the plug is found from Equation (17) to be 3.7×10^{-15} cm². If, as proposed by Millington and Quirk (1959) in dealing with these data, an immobile monolayer has to be allowed for, the calculated value becomes 1.4×10^{-15} cm². These values are of the same order as that of 3.1×10^{-15} cm² obtained by Carman (1952) using liquid freon for the measurement of permeability of another (but similar) plug. This agreement is in keeping with the evidence discussed in Section 2 (IV) on the use of the Kelvin equation for calculating pore size of non-shrinking material from sorption data.

Millington and Quirk (1959) have proposed a permeability equation in which $\varepsilon^{4/3}$ replaces the term ε^2 in Equation (17) and n in unsaturated materials has a different significance. They have used the data of curves 1a and 1b in Fig. 11 and data for Linde silica to calculate permeability and have obtained good agreement with measured permeability. Comparisons for the other materials of Fig. 11 are not given but further details of their equation are to be published.

At the present stage of development of permeability equations it may be said that permeability can be calculated approximately from size distribution of pores without recourse to matching factors or any special parameters relating to other properties. Measured and calculated values have so far been found to agree over a range of permeability from 10^{-15} to 10^{-7} cm² to within a ratio of about 2.3 to one another. Greater accuracy could perhaps be obtained for particular sorts of materials by taking other properties such as particle shape into account but many limitations to accuracy are likely to remain in any general equation.

Permeability equations are of concern to workers on soils, petroleum and water-bearing rocks, filters, and many industrial products. In the field of soil science, they are most needed for dealing with water moving in unsaturated soil where permeability is difficult to determine by direct measurement. This is a similar problem to that in the petroleum industry of calculating permeability for each of two fluids occurring together in porous rocks. A good deal of work is being done in both fields on this subject.

TABLE 8

Intrinsic permeability, k′, of materials represented in Fig. 11

Curve No. in Fig. 11	*Material*	*Fluid; and volume fraction occupied by it*	*Intrinsic permeability, cm²* *Calculated*	*Measured*	*Reference*
1a	Sand 1 to ½ mm	Water $c=0.35$	7.2×10^{-7}	13.1×10^{-7}	Childs and Collis George (1950)
		$c=0.25$	3.2×10^{-7}	4.4×10^{-7}	
		$c=0.10$	0.2×10^{-7}	0.2×10^{-7}	
1b	Sand ½ to ¼ mm	Water $c=0.37$	3.3×10^{-7}	7.5×10^{-7}	Childs and Collis George (1950)
		$c=0.21$	7.6×10^{-8}	8.1×10^{-8}	
2a	Sand 0.8 to 0.4 mm	Water $c=0.30$	4.2×10^{-7}	4.1×10^{-7}	Baver (1938)
2b	Sand 0.4 to 0.25 mm	Water $c=0.30$	2.3×10^{-7}	2.3×10^{-7}	Baver (1938)
3	Oso Flaco fine sand	Water $c=0.42$	2.6×10^{-7}	1.6×10^{-7}	Day and Luthin (1956)
		$c=0.33$	1.2×10^{-7}	1.0×10^{-7}	
		$c=0.15$	1.1×10^{-8}	0.9×10^{-8}	
4	Slate dust	Water $c=0.48$	3.5×10^{-8}	5.0×10^{-8}	Childs and Collis George (1950)
		$c=0.43$	2.8×10^{-8}	3×10^{-8}	
5a	Frio sandstone	Air $\epsilon=0.26$	1.8×10^{-8}	1.7×10^{-8}	Purcell (1949), Fig. 3
5b	Paluxy sandstone	Air $\epsilon=0.20$	8.0×10^{-9}	11.5×10^{-9}	Purcell (1949), Fig. 5
5c	Upper Wilcox sandstone	Air $\epsilon=0.22$	2.7×10^{-9}	4.3×10^{-9}	Purcell (1949), Fig 6
6	San Andres limestone	Air $\epsilon=0.19$	5.0×10^{-10}	3.5×10^{-10}	Purcell (1949), Fig. 7
7	Bradford sandstone	Air $\epsilon=0.15$	2.0×10^{-11}	2.7×10^{-11}	Wyllie and Spangler (1952)

6. DISTRIBUTION OF WATER IN SOIL

1. Application of Darcy's Law

The amount of water in soil is continually changing under field conditions. Water entering from rain or irrigation is distributed fairly rapidly at first, then more and more slowly after the supply has ceased, and is finally removed by drainage, evaporation, or by plants. These processes of gaining or losing water involve movement through soil as a liquid or as a vapour or as both.

Generally the main means of movement through soil is the one described in the chapters on permeability in which water moves as a liquid under the combined influence of gravity and suction. Here Darcy's law (Equation (9) of Section 4 1) applies and can be used for both saturated and unsaturated conditions. In the case of saturated soil, permeability data can be used for example in calculating the rate of flow into artificial drains and generally in analyses concerned with the design of drainage systems. For such purposes saturated permeability is first measured by field or other methods as already discussed.

In unsaturated soil, movement of water at any position in the soil can (and usually does) result in a change in the water content at that position. Darcy's law is difficult to apply to this situation because permeability will change with water content and cannot be treated as a constant. For many purposes it is desirable to know how the water content or suction of a soil will change with time at various positions. If this can be done from a knowledge of initial and boundary conditions and from known properties of the soil, the results will be of value in practical cases concerned with infiltration, evaporation, transpiration, and the accumulation or loss of water under engineering structures.

Some progress towards this objective has been made by combining Darcy's law with the continuity equation. The change in water content with time at a given position is equal to the difference between the rate of flow into and out of the soil at that location or, for flow in the z direction,

$$\partial c/\partial t = -\partial v/\partial z \qquad \text{- - - - - - - - - (19)}$$

where c is the water content by volume and t is the time. Also, since

$$v = -k\,\partial\Phi/\partial z \text{ (Darcy's law)},$$

$$\partial c/\partial t = \partial/\partial z \quad (k\,\partial\Phi/\partial z) \qquad \text{- - - - - - - - - (20)}$$

where k is the hydraulic conductivity and Φ is the potential in cm. For flow vertically, taking Φ as the sum of the capillary potential ψ and the gravitational potential, Equation (20) becomes

$$\partial c/\partial t = \partial/\partial z \quad (k\ \partial\psi/\partial z) + \partial k/\partial z \qquad \text{- - - - - - - - - (21)}$$

If flow is in a horizontal direction the last term (covering the effect of gravity) is dropped. These equations are subject to the effects of previous history of wetting and drying because of hysteresis in the relation between k and c and between c and ψ.

Childs and Collis-George (1950) and Klute (1952) have distinguished the product of k and $\partial\psi/\partial c$ as a diffusion coefficient or diffusivity, D, and on using this term, Equation (21) becomes

$$\partial c/\partial t = \partial/\partial z \quad (D\,\partial c/\partial z) + \partial k/\partial z \qquad \text{- - - - - - - - - (22)}$$

D is somewhat analogous to thermal diffusivity, since it includes a conductivity and a capacity term, but it differs because, when water flows in unsaturated soil, k is not a constant but is a function of c or ψ. The inverse of $\partial\psi/\partial c$ was referred to as the capillary capacity by Richards (1931) in his earlier work with Darcy's law and the continuity equation (19). Kirkham and Feng (1949) have demonstrated that D cannot be regarded generally as constant.

Analysis of water distribution by these equations is not possible for general cases of water entry or drainage of unsaturated soil. Resort has therefore been made to numerical solutions to cover particular cases of water movement. In this way Klute (1952) was able to show that Equation (22) would provide for the wetting front known to occur during entry of water. Numerical applications are difficult to make because of the amount and type of data required and because calculations are tedious. The use of calculated values of k instead of measured values can simplify the data requirements because direct measurement of k is difficult in unsaturated soil. Equation (15) or (17) can be used to obtain k from the curve relating water content and suction for suitable materials. Attention has been given to methods by which the time required for calculation can be shortened (Philip, 1955). However, because of the difficulties, very few comparisons of calculated and measured results have yet been made. Klute, Bruce and Russell (1956) compared the calculated and measured advance of water horizontally in a bed of beads. Day and Luthin (1956) measured the suction (and hence, indirectly, the water content) at various depths and times during the drainage of a column of sand and compared their results with those obtained by calculation. Youngs (1957) measured and calculated the water distribution at various times during the vertical entry of water into slate dust and glass beads. These comparisons of calculated and measured water distribution are encouraging but they are still limited in number and in variety of the materials used.

Gardner (1958) has provided a transformation of Equations (21) and (22) and has shown how this can be used in the solution of steady-state problems such as evaporation of water from a soil in which there is a water table. In his analysis he made use of the relation between permeability and suction which he had found (Gardner, 1956) to fit experimental data obtained by outflow from the pressure membrane apparatus.

Philip (1955b) has shown that vapour movement can be included, as well as liquid movement, in Equation (22). This requires only a redefinition of D, the remainder of the equation being unaltered.

II. Water Entry into Soil

When water is applied to the soil surface it moves downward in the manner described by Bodman and Colman (1944), for a uniform profile, and by Colman and Bodman (1944), for non-uniform profiles. These authors found that the wetted portion of a column of uniform soil (into which water was entering and advancing downwards from the surface) could be considered as being made up of a transmitting zone through which water is transmitted with little change in water content to a wetting zone further down where water content changes markedly with depth and time. At the bottom of the wetting zone there is a sharply defined wetting front advancing downward into the unwetted soil.

The transmitting zone was found by Bodman and Colman to be an ever-lengthening, unsaturated zone of fairly uniform water content and suction except that at the top there is a saturated and swollen condition just below the surface on which the water is ponded. The common characteristics of the transmitting zone (except for the part near the surface) appear to be saturation to the extent of about 0·8 of the pore space and suctions of less than 25 cm.

Miller and Richard (1952) observed that suction increased with depth within the transmitting zone but that the gradient in that direction decreased as the zone lengthened with time. For their work, they developed a tensiometer system designed for use where the suction is changing rapidly and they were thus able to make their measurements at early stages in the infiltration process. The measurements of Marshall and Stirk (1949) made after water had penetrated fairly deeply ($>$70 cm) showed a small gradient to be present in the transmitting zone of a clay column but not in a sand and a loam column. The results of Taylor and Heuser (1953) also showed that the suction gradient approached closely to zero by the time the water had penetrated to that depth in their cores. Hence, after water has penetrated sufficiently deeply into the soil, movement within the transmitting zone is due principally to gravity. These characteristics are little altered when the rate of entry is restricted somewhat as by a preceding horizon of lower permeability (Colman and Bodman, 1944) or by spraying the water on to the surface at a lower rate than that at which it would enter if ponded on the same soil (Marshall and Stirk, 1949).

If, on the other hand, a horizon of low permeability lies beneath one of high permeability on which water is ponded, positive pressures will develop in the top horizon in a confined column of soil (Colman and Bodman, 1944).

The infiltration rate, determined by applying excess water to soil, decreases with time from the commencement of the trial; and the rate of decrease also decreases with time as shown by the widely observed tendency for infiltration rate to approach a final minimum value. For practical purposes infiltration results are often presented simply as the amount entering in a given time, or as the time required for a given amount to enter. The full results for a particular trial can be plotted as a cumulative curve representing the total amount, i cm^3/cm^2, which has entered in t secs. The values of the parameters in empirical equations fitting these field data more or less satisfactorily are frequently used to represent the results. Two different equations are used for this purpose. One of these

$$i = a\,t^b \qquad \text{- - - - - - - - - (23)}$$

is due to Lewis (Anon, 1926; Lewis, 1937) and Kostiakov (1932) and is usefully applied to the early stages of the infiltration process. The two parameters a and b can be evaluated for each trial by fitting a straight line to the data when log i is plotted against log t. The second equation is due to Gardner and Widstoe (1921) and to Horton (1940). It has three parameters one of which represents the fairly steady rate reached after water has been applied continuously for a considerable time. This "minimum infiltration rate" has been much used in watershed studies of infiltration and run off. Under these conditions Equation (23) may fail in that, with $b<1$, a zero rate of infiltration is implied at high values of t. On the other hand in the case of irrigation land, where water applications are not long continued, Equation (23) is appropriate and is more commonly used.

Kirkham and Feng (1949) have reviewed experimental evidence which shows that when water penetrates horizontally into a porous medium

$$i = c\,t^{\frac{1}{2}} + d \qquad \text{- - - - - - - - - (24)}$$

where c and d are constants, d being of small magnitude. This equation applies in the absence of any effect of gravity upon water entry and hence in the case of infiltration vertically, b in Equation (23) should be about $\frac{1}{2}$ when gravity is relatively unimportant. This will be the case when there is a high suction gradient affecting entry as in the early stages of advance into soils which are initially dry and into soils of high clay content. In wet and sandy soils, where the effect of gravity is relatively more important, i tends to be more nearly proportional to t, and b is more nearly 1.

The two empirical infiltration equations of Lewis and Horton have been compared and discussed by Philip (1957) in one of a series of papers concerned with the theory of infiltration. Methods for solving the diffusion equation (22) for the case of infiltration have been developed by Philip who has derived the following simplified equation from his analysis,

$$i = S\,t^{\frac{1}{2}} + A\,t \qquad \text{- - - - - - - - - (25)}$$

Here S is the "sorptivity" (a measure of the uptake of water by the soil) and A is a parameter which arises from the analysis.

Infiltration measurements are commonly made in the field by ponding water on the surface or by spraying it on at a rate great enough to cause water to run off the surface into a measuring vessel. In either case the amount of water in cm or inches entering the soil can be recorded as a function of time. A ponding method is described by Richards (1954) and different types of spray methods have been compared by Wilm (1941). If large plots are difficult to handle because water is not available in quantity, a convenient method is to use a metal ring about 30 cm in diameter inserted a few cm into the soil and containing water ponded to a small constant depth from a Mariotte tube leading from a calibrated drum. Small test areas of this sort are, however, subject to the complication that much of the water spreads laterally beyond the limits of the area covered by the water. This will not affect results seriously if the permeability of the surface soil is limiting (Slater, 1957). If it is not limiting the rate of infiltration will be much greater than that which would be obtained using a larger wetted area and the rate as measured will not be independent of the size of plot used in the test. Various ways of reducing this error have been

tried and some of them have been discussed and examined by Marshall and Stirk (1950). Wetted buffer areas surrounding the test area are generally effective when the water is sprayed on (Duley and Domingo, 1943) but are not always very satisfactory for ponded water. Deeply penetrating cylinders (Musgrave, 1942) are better than shallow rings if they can be inserted without disturbing the soil unduly. Prior wetting of the site and rejection of extreme results largely remove the error in deep uniform soil (Burgey and Luthin, 1956). When the surface is not limiting an empirical correction can be applied to the infiltration data (Kachinsky, 1934) using for this purpose the fraction of the applied water which is found to have penetrated vertically when the soil is examined at the end of the test. As shown by Marshall and Stirk (1950), results from small plots are then reduced to values which are of the order found for large plots and at the same time the variability caused by erratic spreading is also reduced. The correction is only a rough one since there is no check on the distribution of water at intermediate times during the infiltration trial.

All those soil and water conditions which affect permeability can also affect the rate at which water enters the soil. The rate is much affected by the initial water content of the soil (Tisdall, 1951). When this is high the potential gradient in the wetting zone and also the size of the channels and pores in swelling soils will be reduced and there will consequently be some reduction in the infiltration rate. Thorough drying can greatly increase infiltration rate (Reitemeier *et al*, 1948) and, as discussed in Section 3 II, the effect may last for some time. A surface skin of low permeability will sometimes develop and control the rate of infiltration even though it may be only a few mm thick (Duley and Kelly, 1939) and the effect of this skin on infiltration may change progressively while water is in the process of entering the soil (McIntyre, 1958). The extent to which structure is altered depends upon the intensity of rain and the protection afforded by plants and mulches of straw and other materials as well as upon the properties of the soil itself. The infiltration characteristics of soils are thus not easily assessed—they vary with the season, the condition of the surface soil, the type of vegetation, and the way in which the water is applied.

Under these circumstances it is obviously difficult to establish any standard relations between infiltration rates and soil conditions. Musgrave (1955) suggests tentatively that soils can be divided into the four groups shown in Table 9 according to their minimum infiltration rate as measured under minimum cover after a long and excessive rain. These are rather exacting conditions and many published rates are much higher than those given in the table.

TABLE 9

Minimum infiltration rate of various soils arranged in four groups (From Musgrave, 1955)

Minimum infiltration rate, in/hr	*Soils*
0·45-0·3	Deep sands, deep loess, aggregated silts
0·3-0·15	Shallow loess, sandy loams
0·15-0·05	Many clay loams, shallow sandy loams, soils low in organic matter, soils of high clay content
<0·05	Soils of high swelling, sodic soils

Infiltration measurements are used for testing soils in use or proposed for use under irrigation, for examining the hydrology of watersheds and as a means of finding out the effects of various practices and soil treatments on infiltration. The methods to be used and the interpretation of the results obtained are matters requiring judgment. In irrigation studies ponding methods are commonly used unless a check has to be made on the suitability or effects of sprinkler irrigation. In watershed problems spray types of infiltrometer are preferable to flood types because of the closer resemblance to rain action on the surface. Infiltration is also studied on the basis of stream gaugings of run-off during storms of measured rainfall instead of by small plot methods when this can be done.

III. Drainage and the Field Capacity of Soil

Soil which has been fully wetted down to a depth at which water is escaping at atmospheric pressure by means of some drainage outlet, will go on draining until the gradient in the total potential is zero (i.e. the soil at a height z above the drain is at a suction of z cm). During the approach towards this condition, the rate of water movement decreases with time partly because of decreasing permeability but also because the suction gradient opposing flow increases. This condition was observed by Luthin and Miller (1953) in soil columns 122 cm long after draining for 1 or 2 days into the atmosphere. The permeability of the soil a short distance above the level at which the water is at atmospheric pressure (the water table) will not decrease very much during draining if the suction is not sufficient to empty much of the pore space. As noted by Luthin and Miller, the difficulty of draining clay soils may be due more to the relatively high suction required to get air to enter the small pores than to low permeability. The zone just above the water table which thus remains almost saturated after draining is referred to as the capillary fringe. However, in soils with widely distributed pore sizes, the concept may be rather artificial because both the extent and degree of definition of the capillary fringe depend upon the size distribution of pore space.

A number of methods are in use for draining soil artificially so as to lower the water table and remove excess water from soils which are too wet. These include open ditch, tile, and mole drain systems. The need for drainage may be judged from appearances or from records of water table fluctuations. The depth and spacing of drains for a particular environment and cropping system depend greatly on the permeability of the soil horizons. As noted earlier, this property is usually examined in an indirect and qualitative way by subsoil inspection. For straightforward cases, backed by good local experience, this is usually sufficient. For unusual or difficult problems, actual measurement of permeability is preferable. When such data are available numerical solutions can be reached for particular cases. In addition a number of general solutions of certain drainage problems have been worked out. A comprehensive review of these and other aspects of drainage research is given in "Drainage of Agricultural Lands" edited by Luthin (1957).

Water applied to dry soil can go on moving downwards long after the application has ceased. Edlefsen and Bodman (1941) found appreciable downward flow (at a rate of 0.007 in/day) still occurring after 590 days at a depth of 9 feet in a silt loam soil covered against evaporation. There is thus no early end to downward movement into dry soil. After the supply has been cut off at the surface, the transmitting zone becomes a draining zone in which the water content decreases and the suction increases with time. It follows from Equation (18) that permeability also decreases progressively with time. Also, as shown in Fig. 12, the suction gradient which can favour downward flow in the transmitting zone, later opposes flow after the supply at the surface ceases (Marshall and Stirk, 1949). For both these reasons of changing permeability and changing suction gradient (but especially permeability) the rates of movement in all parts of the draining zone are very much lowered after a day or two. The water content of the draining zone will subsequently change only slowly and the soil is said to be at field capacity.

In agriculture, the water content at field capacity is significant as representing the amount of water which stays long enough in a particular horizon or profile to be of possible use to plants subsequently. The water content at this stage is a fairly reproducible and useful characteristic of soil in the field provided a standardized field procedure is followed in determining it so as to reduce the variable effect of boundary conditions outside the sampling zone. It is a more clearly marked stage in sandy soils than in clays because the large pores of the sand soon empty and permeability decreases rapidly as a consequence. Moore (1939) noted that although the order of permeability of soils may be sand>loam>clay for suctions lower than 100 cm, the order will be reversed for suctions greater than 100 cm. There will thus be a sharp change in permeability of most sands but only a gradual change in permeability of many clays as drainage proceeds and this will account for differences in drainage behaviour at field capacity.

Field capacity is measured by sampling the soil a day or two after it has had a thorough wetting. The horizons to be sampled need to have been fully wetted in the first instance and for this reason it is advisable to add enough water for the wetting front to have penetrated 30 cm or more beyond the proposed maximum sampling depth (Colman, 1944). The soil needs to be kept free from appreciable losses of water due to evaporation or transpiration. The water content

at field capacity is most usefully expressed on a volume basis (cc water per cc soil). The depth of water in a given thickness of soil horizon can then be calculated readily. On summing the depth of water in each of the successive horizons concerned with plants and their water requirements, the field capacity of the profile to a stated depth is obtained.

Results obtained from various types of laboratory measurements on samples have been found to correspond fairly well in certain cases to water content at field capacity. On samples whose structure has been partly destroyed, the moisture equivalent and the one-third atmosphere percentage are used—the first being the water content by weight after centrifuging in a specific manner and the second being the water content by weight after a suction of ⅓ atmosphere has been applied. These methods are simple and rapid but, if only because of altered structure, there are bound to be exceptional values which are not in good agreement with field capacity. Thus in sands the moisture equivalent is usually very much lower than field capacity and in clays the reverse may hold. There are obvious advantages in working instead with samples whose structure has been preserved and determining their water content when a suction typical of the draining zone at field capacity has been applied. There is, however, no particular suction which can be said to correspond accurately to field capacity. Suction decreases with depth and increases with time in the draining zone of a given soil and is often of a lower order in clays than in other soils. The reported values at field capacity mostly lie between 50 and 100 cm, but no laboratory method can be a real substitute for a field capacity measurement which is necessarily influenced by many other factors including especially the physical properties and initial moisture conditions of the profile as a whole. Thus a subsoil horizon of low permeability will delay the drainage of the overlying horizons causing them to have a higher water content at the end of a two-day period than they would have if drainage were unrestricted. This will happen also if the subsoil is initially very wet due to a shallow water table. Field capacity data reflect all such field conditions whether they are permanent or temporary features of the profile. However, for certain purposes,

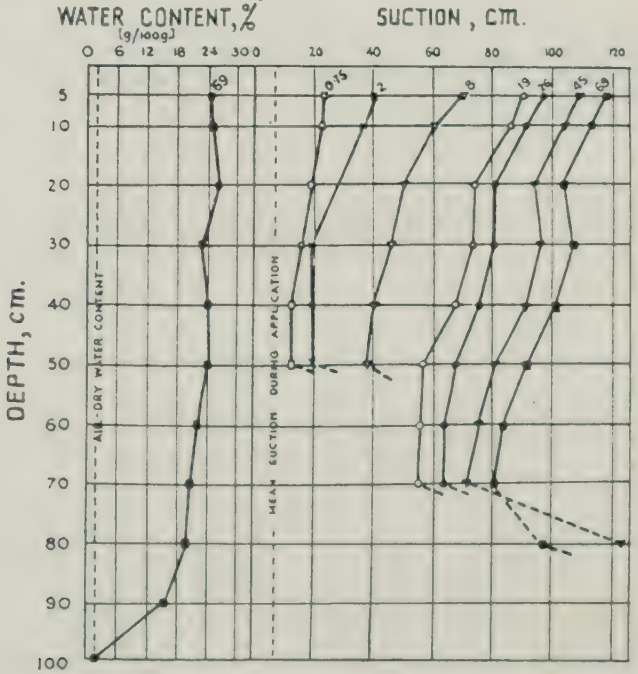

FIG. 12. *Water content and suction during drainage of a column of loam soil. The period of draining is given in hours (Marshall and Stirk, 1949)*

a suction method is often preferred since it gives more precise and more permanently useful information about each horizon relating to the sizes of the pores (and hence to water and air movement generally) as well as to water retention at field capacity. The water retained by naturally structured soil at 100 cm suction is often measured for this reason.

The suction in sand, loam and clay columns and in field soils for water draining into dry soil (after the supply at the surface has ceased) has been examined by Marshall and Stirk (1949). Some of the data for their loam column are given in Fig. 12 which shows how suction can vary with time and depth in the drainage zone. The development of a suction gradient opposing downward flow is also illustrated by the data. One additional feature of the moisture profile at field capacity can be seen from the curves—there is an apparent disparity between the water content and suction data since values for both of these decrease with depth in the drainage zone. This anomalous result is substantiated by other data discussed by the authors and is evidently a feature of the moisture profile of many soils at field capacity—more especially if they are of medium or high clay content. Schofield (1935) had earlier stated this disparity was likely to occur. It may be a consequence of hysteresis and of the increased restraint on swelling at increasing depth. The effect was very marked in the clay column where swelling occurred; but there appears also to have been some effect in the lower part of the sand column where it did not. In the light of their work on structure, Emerson and Grundy (1954) attributed it to the more rapid wetting near the top of the columns. This (as is discussed later) causes greater aggregate disruption and greater water retention as a consequence of the disruption.

7. MOVEMENT OF WATER VAPOUR

I. Introduction

Water enters into soil and is removed by natural or artificial drainage and by plants mainly by the process of flow in the liquid phase under essentially isothermal conditions. However, movements can occur in the vapour as well as in the liquid phase and under the influence of a temperature gradient as well as under isothermal conditions. In particular, seasonal and diurnal changes in temperature may in this way be associated with evaporation of water from soil, slow moisture changes under roads and buildings, and water accumulation near the surface in the winter in cold regions.

Vapour may move because of pressure gradients set up in the soil air by wind passing over the surface, but Fukuda (1955) found wind gustiness to be rather a minor cause of vapour transfer. Penetration of the wind was to a depth of only about 5 mm. This work was on rather finely structured soil and experiments by Greacen and Gurr (1958) show that evaporation of water from rough fallows and from soil with cracks opening to the surface may be influenced a good deal by wind. Their experiments show that either wind or radiant heat can cause convective movement of air in the soil and that the short term losses of water from soil of cloddy structure are much greater than those from fine. They believe that this may be important in regard to conserving water and mobilizing plant nutrients during fallow periods under certain circumstances. Their results on structure and evaporation are in general agreement with those outlined in Section 1 (III). The optimum size of aggregate required to bring the rate of evaporation to a minimum under the conditions of this experiment was about the same as that found by Burov (1954).

Water can move as a vapour through soil by diffusion in the direction of decreasing vapour pressure. A gradient in vapour pressure can be caused by gradients within the soil in salt concentration, suction, or temperature. The effect of dissolved salts is to depress the vapour pressure of the soil solution and thus cause a gradient of vapour pressure which depends on the gradient in concentration. The effect of suction on vapour pressure is given in Equation (2) of Section 2 (II). From this it can be shown that, in moist soil, a large gradient in suction will cause only a small gradient in vapour pressure. (For example, in passing from saturation to water at 15 atmosphere suction, the vapour pressure relative to that of free water drops only from 1 to 0·98.) Hence within this range vapour movements from this cause are negligible. Only if there is a gradient from moist to dry soil conditions will movement as a vapour be appreciable under isothermal conditions. Since the relative vapour pressure of "air-dry" soil may be below 0·5, a large gradient in vapour pressure can occur between wet and dry parts of a soil and appreciable vapour movement has been demonstrated over short distances in these circumstances. The third cause of a gradient in vapour pressure—the existence of a temperature gradient—is generally the most important of the three. Temperature has a great effect on vapour pressure of water which, for example, is more than tripled in changing from 10° to 30°C.

II. Diffusion Equations

The quantity, q, of a gas or vapour diffusing through a mixture with other gases across unit cross-sectional area in unit time under a partial pressure gradient dp/dl, is given according to Fick's law by

$$q = - D_o \, dp/dl$$

where D_o is the coefficient of diffusion. The introduction of a porous medium with a gas-filled porosity of ε_g reduces the area through which gas can diffuse to a fraction ε_g, of the cross-

sectional area of the medium. The diffusive flow, q_m, of the gas across unit cross-sectional area of the medium then becomes

$$q_m = -D_o\, a\, \varepsilon_g\, d\,p/d\,l$$

where a is a factor which allows for the fact that rate of movement through the pores cut by any plane is reduced by the poor connections they make with other pores in the apparent direction of movement normal to that plane. If D is the diffusion coefficient of the gas in the porous medium (so that $q_m = -D\ dp/dl$), then $D/D_o = a\,\varepsilon_g$. Penman (1940), following Carman's tortuosity model (Section 5 II), treated the obstructions as though they caused a lengthened path and hence a reduced partial pressure gradient in the pores. This leads to a constant value for a of $1/\sqrt{2}$ and Penman adopted a value 0.66 close to this as giving the best fit to his data. Hence the equation which Penman obtained and which has since been much used in problems of aeration and movement of water vapour is,

$$D/D_o = 0.66\,\varepsilon_g \qquad \text{- - - - - - - - - (26)}$$

However, the chance that a pore will communicate effectively with other pores should increase with porosity and it is thus unlikely that a is really constant. The fraction, ε_g, of unit area in a plane normal to the direction of movement will behave more as a completely free area for diffusive transfer of gas the more closely ε_g approaches 1 because of improving interconnections. Since the pore area represented by ε_g is not a completely free area, a can be considered as a hypothetical fraction of this pore area which is free from obstruction, the remainder of the pore area being regarded as inoperative. The fraction, a, was found by Marshall (1959) to be given by $\varepsilon_g^{\frac{1}{2}}$ from a consideration of Equation (17) for permeability of a porous medium, and the expression $D/D_o = a\,\varepsilon_g$ can in this case be written as

$$D/D_o = \varepsilon_g^{3/2} \qquad \text{- - - - - - - - - (27)}$$

The step taken in Equation (27) allows for improved connections between pores with increased porosity. This was done arbitrarily using a permeability model; and agreement obtained with experiment is only approximate as is shown, for the data of Taylor (1949), in Fig. 3 on p.4. Using a quite different approach, the ratio of the electrical conductivity of a material saturated with brine to that of the brine itself can be taken to be analogous to D/D_o. The inverse of this ratio is much used as the "formation resistivity factor" of the oil industry. For this ratio, Wyllie and Gregory (1953) found that an expression of the form $c\varepsilon^k$ (where c and k are constants) would fit their data on cemented materials and that expressions like (27) of the form ε^k were not adequate.

III. Mechanisms of Water Movement under the Influence of a Temperature Gradient

In soil which is moist but not very wet, water is found to move in the direction hot to cold (i.e., from high to low vapour pressure) at a rate which is greater than that to be expected from Fick's law if the process is regarded as one of diffusion of water vapour (Smith 1943, Gurr, Marshall and Hutton, 1952; Rollins, Spangler and Kirkham 1954; Taylor and Cavazza, 1954). In the neighbourhood of permanent wilting point it may be 4 or 5 times too great. The following explanations and mechanisms have been advanced for this.

a. Thermo-osmosis. Bouyoucos (1915) and Winterkorn (1947) suggest that water in films around the particles moves because of a change of water affinity with temperature. Water affinity is said to be related to the electrical double layer and hence to the amount and type of the exchangeable cations. Thermo-osmosis is considered by Winterkorn to be similar in principle to electro-osmosis (the movement of water under the influence of an applied electrical potential). Deryaguin and Melnikova (1956) consider that thermo-osmosis cannot be of importance in soils of low water content and the small flows which they obtained were in the reverse direction cold to hot. Under saturated conditions the effect may be greater and it is to this wet condition that the allied electro-osmosis techniques have been applied with some success to assist drainage (Casagrande, 1952).

b. Thermo-capillary movement. Deryaguin and Melnikova (1956) explain the large movements at low water contents as due to the effect of temperature on surface tension of water which decreases with increasing temperature. It will be seen from Equation (1) of Section 2 (II), that

suction will therefore decrease with temperature and there will thus be a suction gradient capable of moving water toward the colder part of the soil. Movement then occurs in accordance with Darcy's law in that direction. Deryaguin and Melnikova quote Rozhanskaya and Godun as having found good agreement between flow toward the surface soil (as measured from water content changes) and flow as calculated on the basis of thermo-capillary transfer in winter under field conditions near Leningrad. Flow amounted to 10 cm of water from 16th December 1953, to 26th February 1954, and the authors consider that these movements could not have been accounted for by vapour diffusion. The phenomenon of surface accumulation of water in cold winters in the south-east of the European part of the U.S.S.R. is believed by the authors to be due to this mechanism. Lebedeff (1927) has also measured water accumulated in surface soil at Odessa and has ascribed this to the effect of a temperature gradient.

c. Vapour movement. The failure of Fick's law for diffusion of gases to account fully for the movement of water under a temperature gradient was explained by Smith (1943) as being possibly due to the rapid transfer of water when vapour condensed on one side and evaporated on the other side of small liquid bodies in the porous medium. Philip and de Vries (1957) have given an expression for movement using a model somewhat similar to Smith's and allowing also for different temperature gradients in the different phases. They consider that small isolated rings of water receive the vapour and transmit it, at the same rate as it is received, to the next pore. The water transfer on this basis will be greater than by simple vapour diffusion alone (Equation (26)) and the ratio between the two was taken as

$$\eta = \frac{\varepsilon_g + c}{0{\cdot}66\,\varepsilon_g} \quad \zeta$$

where c is the water content by volume and ζ is a variable which is intended to allow for a higher temperature gradient in the pores than in the soil as a whole. The ratio, η, was found to bring about good agreement between calculated and measured transfer in the data of Gurr, Marshall and Hutton (1952) and of Rollins, Spangler and Kirkham (1954). A difficulty in this treatment is that the liquid phase is assumed to be discontinuous and an arbitrary factor has to be introduced to meet the cases where this cannot be true.

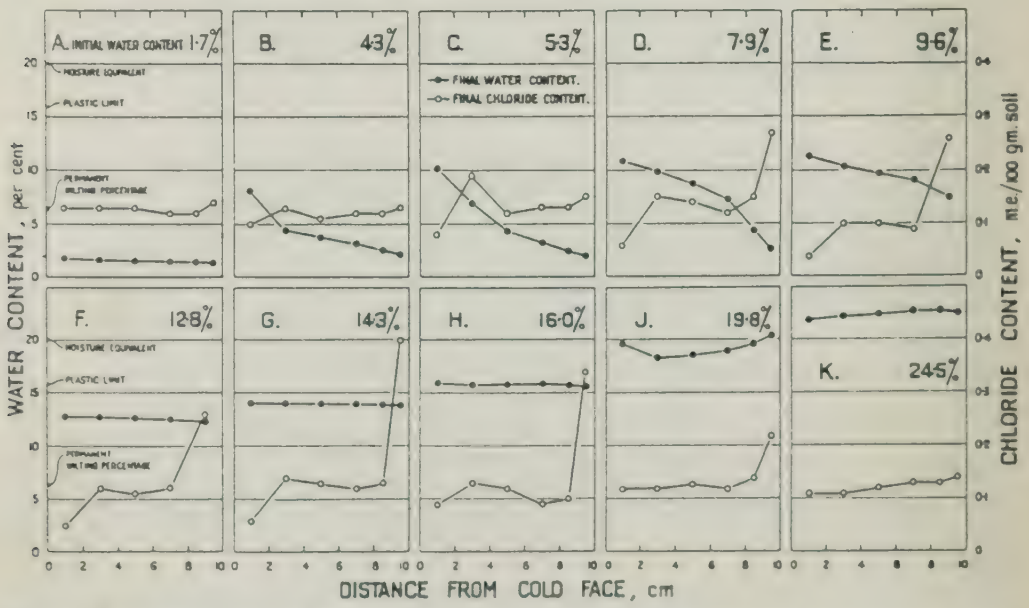

FIG. 13. *Distribution of water and chlorides in columns of a loam soil after being subjected to a temperature gradient for five days (Gurr, Marshall and Hutton, 1952)*

iv. Evidence of Diffusion under the Influence of a Temperature Gradient

Gurr, Marshall and Hutton (1952) have examined the contribution from vapour transfer and thermo-capillary transfer of liquid water in an enclosed soil column of uniform water content to which a temperature gradient was applied. Some of these results are shown in Fig. 13. By measuring the transfer of water and also of solutes they were able to distinguish water movement that occurred wholly in the liquid phase. They found a nett transfer of water towards the cold end (except for their wettest columns—Fig. 13K) but the solutes moved in the reverse direction. Hence if any liquid moved towards the cold end at all due to thermo-capillary flow, this process must have been soon stopped by the development of a moisture and suction gradient favouring liquid movement from cold to hot soil. It is evident in Fig. 13F G and H, that only a minor change in moisture distribution is required to cause this reversal of the suction gradient. A circulation of water was thus established in which vapour went towards the cold end and liquid towards the hot where the salt was found to accumulate progressively. Rollins, Spangler and Kirkham (1954) conducted a similar experiment but in their case the return liquid flow from cold to hot took place in an external tube and not in the soil itself. They were thus able, at the start of each run, to look for evidence of thermo-capillary flow from hot to cold in the external tube but they detected none. Both groups of workers found the vapour movement to be about four or five times greater than that expected by diffusion according to Fick's law and this was no doubt due to the interposition of liquid films as already discussed.

Other workers have also found that a circulation of liquid and vapour occurs in closed systems of various porous media under the influence of a temperature gradient (Kirscher and Rohnalter, 1940; Hutcheon and Paxton, 1952; Taylor and Cavazza, 1954; Hadley and Eisenstadt, 1955; Kuzmak and Serada, 1957) and have ascribed the transfer of water toward the cold end to vapour movement. Using a different approach, Jennings, Heymann and Wolpert (1952) found that air pressure affected the rate of transfer to the cold end and took this as evidence that movement occurred by diffusion. Jennings *et al* and Kuzmak and Serada also measured electrical potential during movement under a temperature gradient but did not consider this had a significant role in the movements observed. Thus the evidence in support of vapour diffusion as the basis of movement under a temperature gradient is substantial provided some influence of the liquid phase is also admitted to account for the high rates of movement observed. Thermo-osmosis and thermo-capillary movement may occur in particular circumstances, but the most usual mechanism appears to be basically one of vapour diffusion.

The rate of movement towards the cold end is dependent on the temperature gradient (*see* Rollins *et al*), the water content, and the air-filled porosity. A maximum rate occurs in fairly dry soil at about the permanent wilting point (Fig. 13C and D). It is lower if the soil is too dry (Fig. 13A and B) and there is a progressive decrease with increasing water content which can be illustrated for fairly moist soils by the data of Rollins *et al* (1954). They used soil packed at different bulk density but mostly of uniform water content by weight. When changed to a volume basis (as has been done in Table 10), their data are seen to cover a fairly extensive range of water content and of air-filled pore space. Within these ranges, rate of movement increases with air-filled pore space and decreases with water content. The data are for systems in which the movement from hot to cold occurred within the soil and the return liquid flow occurred in an external tube.

All these results on movement under a temperature gradient are from laboratory experiments in which the gradient is maintained despite the heat transfer which accompanies evaporation and condensation of water. In the field, the gradient is not so likely to persist when water transfer occurs and few examples have been studied. In their long-term experiment on downward movement of water, Edlefsen and Bodman (1941) attributed reversal which they found in the direction of movement in a deep subsoil, to seasonal temperature changes. Richards, Gardner and Ogata (1956) noted an effect of temperature gradient near the surface on downward movement and this will be discussed further in the next section.

TABLE 10

Movement of water under a temperature gradient of 3.8°C/cm in a silty loam soil packed to different bulk density

(From data of Rollins, Spangler and Kirkham, 1954)

Trial	*Porosity*	*Air-filled porosity*	*Water content by volume*	*Rate of movement through soil* 10^{-3} *cm/hr*
SL-1	0.69	0.52	0.17	8.10
5 to 8	0.53	0.32	0.21	5.38
SL-2	0.52	0.26	0.26	5.35
1 to 4	0.50	0.20	0.30	4.40
SL-3	0.45	0.15	0.30	3.33
SL-H-1	0.41	0.08	0.33	2.56

8. SALT MOVEMENTS RESULTING FROM GAINS AND LOSSES OF WATER

1. Leaching of Salts

The distribution of readily soluble material in the soil is susceptible to change with season and with agricultural practices. The most upsetting practices are those of irrigation and drainage, but as already noted, changes in plant cover which affect transpiration losses may also affect the movement of water and hence of solutes through the soil. The introduction of irrigation practices to soil which has been little leached during its development, greatly affects the distribution of soluble material. If the soil is permeable, much of this will be quickly taken down below the rooting zone of the irrigated plants. Thus, in the brown solonized soils of the Berri and Cobdogla Irrigation areas in South Australia, the chloride content to a depth of 200 cm was reduced by the introduction of irrigation to less than one-fifth of the content of virgin soil over most of the district. In those parts of the district where salinity has become a problem, this is usually a secondary result of the formation of a local water table from which water moves upward and then evaporates leaving salts near the surface. In these cases, the redistribution of salts due to upward movement of water can result in a concentration of salts within the rooting zone which is far greater than that originally in the virgin soil and this has caused injury to plants. Artificial drainage at a depth of about 200 cm is provided to remedy or prevent this.

Despite the greatest care in controlling irrigation applications, it is hardly possible to irrigate land adequately without adding some water (either from rain or irrigation) to a depth where it is beyond the use of plants. In fact it is often necessary to do this in order to remove from the root zone the salts being added continually in small amounts in the irrigation water. Contributions to the deep zone come also from underground seepage of water from supply channels and from excessive applications to higher land. Sooner or later a water table usually develops as a result of these contributions, and underground drainage is then necessary to prevent salt from accumulating near the surface when water, moving upward, evaporates from wet soil.

In soil of low permeability the removal of salt is often more difficult. For example in certain soils in the Wakool Irrigation District, New South Wales, the salt content has been reduced only to a depth of about 30 cm after several years of irrigation. In this case the clay has appreciable sodium in the exchange complex (a sodic soil) and hence a reduction of permeability accompanies the removal of salts because of dispersion. It may be necessary in such cases to replace sodium by calcium by applying gypsum or other source of calcium in order to maintain the permeability of the soil. A similar problem can arise in land which has been flooded with sea water. These effects on permeability have been discussed in another section. For an account of the reclamation and management of saline and sodic soils reference may be made to a review by Fireman (1957).

The control of salt is mainly a matter of controlling water movement and the principles already discussed on permeability, infiltration and water distribution thus apply to salinity. However, there are some differences of detail. When dealing in Section 4 with flow of water through soil on the basis of Darcy's law, the macroscopic velocity was considered. In fact the actual flow is characterised by changes in the direction and magnitude of the velocity as the water passes from one pore to the next and Scheidegger (1954) has provided a statistical theory for the resulting dispersion which occurs in a fluid moving through a porous medium. On the basis of this theory, Day (1956) has derived equations for the dispersion of a salt-water front advancing through water-saturated porous material. Instead of a sharp front separating the new salt solution from the old water, there is a diffuse zone whose length increases as the square root of the distance travelled. Day (1956) and Day and Forsythe (1957) found this to be the case in sands and soils.

In addition to the effect of distance travelled upon the dispersion of the front, there is also probably an effect of velocity. When a new salt solution enters a pore it will not completely displace the water on a front within the pore. Velocity will be greatest at the centre of the pore and the lower it is the greater will be the opportunity for molecular diffusion of the salt to occur outward from the centre. Taylor (1953) has derived an expression for the dispersion that will consequently occur in a single capillary tube. From this it can be shown that for a given velocity dispersion is proportional to the square root of the distance travelled, and, at a given distance, it is also proportional to the square root of the velocity. This effect of velocity on dispersion is discussed by Scheidegger (1957; 1958). It was not apparent in the results of Day and Forsythe but was observed by Von Rosenberg (1956) in sand. This and other questions on dispersion are being actively studied in the field of oil technology.

Day and Forsythe conclude that the mixing which is known to occur under laboratory conditions during the displacement of soil solution with water is a consequence of dispersion and, under field conditions, dispersion affects the distribution of plant nutrients applied at the surface and the leaching of salts from soil. Other examples are to be found in groundwater hydrology, oil technology and chemical engineering.

II. Upward Movement of Salts

Views on the significance of upward movement of dissolved material have fluctuated over the years. Keen (1931) noted that some early workers considered movement in the liquid phase could take place 10 metres or more above a water table but there was substantial evidence to the contrary and he found 35 cm and 80 cm above a water table to be reasonable practical limits for a coarse sand and a heavy loam respectively under English conditions. This conclusion was later backed by measurements of permeability of unsaturated soils from which it appeared that most soils approached zero permeability at quite small suctions, (Moore, 1939). However, it was recognized that the zero limit was not reached in these experiments, and it is now known that water can move in the liquid phase under very much drier conditions than that (Lewis, 1937; Richards, 1941; Marshall and Gurr, 1954). The movement of salts under relatively dry conditions is demonstrated in Fig. 14 which shows that chlorides were able to move away from the lower to the upper half of unsaturated blocks of soil 2 cm high when water was allowed to evaporate from the top. The soil in each block had a uniform distribution of chlorides and water at the start. After 24 hours, half of the chlorides had moved out of the lower half of the block even when the soil was little wetter than the permanent wilting point. In five other soils also examined, chlorides were found to move under similar conditions.

It follows that movement of solutes can occur under drier conditions than those associated with a shallow water table when circumstances are favourable. If these movements are to be over any considerable distance they will require long periods of time to be of significance. When a favourable gradient is established for some months, there will be an opportunity for appreciable upward movement from water tables at depths at least as great as 250 cm (e.g., Shaw and Smith, 1927) provided the soil has a large proportion of pores of the right size to remain full under these conditions. In soil whose pores can mostly be emptied by applying a suction of h cm to the water, upward movement is limited to heights less than h cm above the water table. For this reason soils with mainly large pores such as sands will have a low limit for upward rise. Thus Jackson, Blackburn and Clarke (1955) found that salt accumulated near the surface of a sandy soil if the water table was not deeper than about 100 cm while Thomas (1939) found that salt accumulated in amounts toxic to plants at the surface of soil of medium texture when the water table was as deep as 200 cm. The limiting heights which might be considered of practical importance depend upon the local climate as well as the soil. These two cases were observed during a summer period of about three months during which there was not enough rain to cause significant downward movement.

Gardner and Fireman (1958) have presented the results shown in Fig. 15 for rate of evaporation from a simulated water table at the stated depths in a sandy loam column. In their theoretical treatment the main movement is shown to be in the liquid rather than the vapour phase and hence Fig. 15 can also be considered as a guide to the upward movement of salt dissolved in the water. For this column it can be seen that the rate of movement of salt towards the surface is seriously

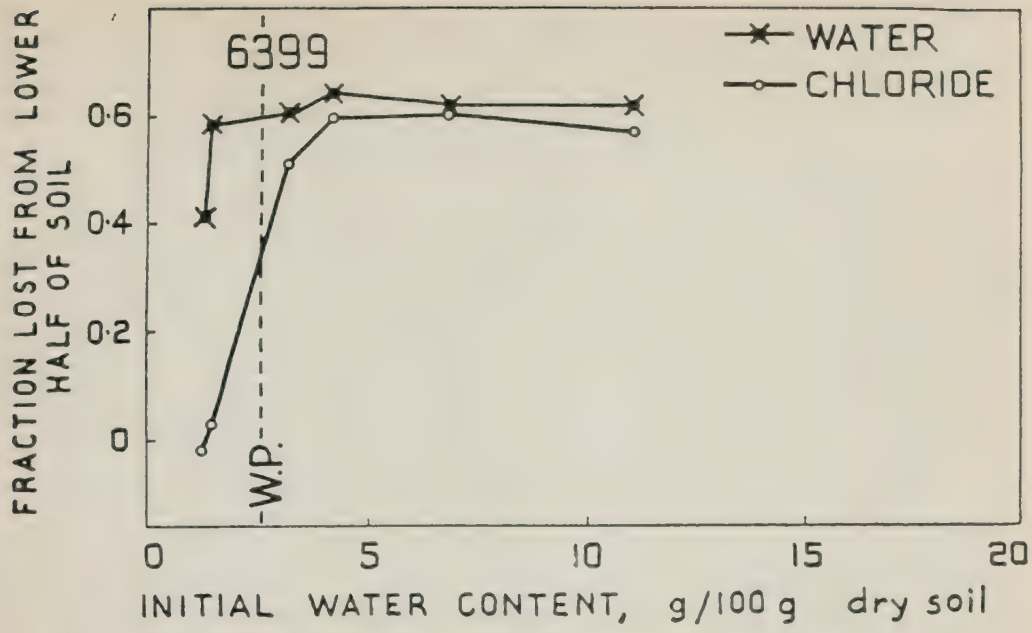

FIG. 14. *Movement of water and chlorides in relatively dry soil from the lower to the upper half of a short (2 cm) column of sandy soil from which water is evaporating. The losses of water and chlorides from the lower half when the soil is at different initial water contents are expressed as fractions of the amounts initially present. W. P. represents permanent wilting point (Marshall and Gurr, 1954)*

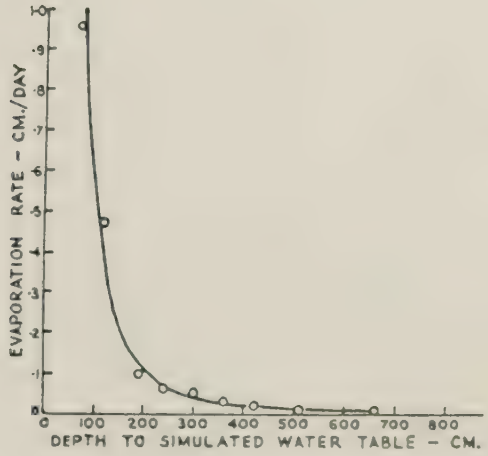

FIG. 15. *Theoretical and experimental rates of evaporation of water from a column of sandy loam soil as affected by depth to simulated water table (Gardner and Fireman, 1958)*

high if the water table is more shallow than 100 cm; that it declines greatly with lowering of the water table to about 200 cm, and that further lowering beyond that depth would not affect the position so greatly. Hence in this case, considering only the lower efficiency of shallow drainage and the higher cost of deep drainage, a depth of about 200 cm for laying the drains is indicated to provide protection from salt. In practice, a number of other factors affecting depth of drainage have to be considered as well and some of them have already been touched on.

It can be seen from Fig. 14 which shows that solutes can move in soil of low water content, that some redistribution of solutes is to be expected upon drying out a soil whether or not there is a water table. Certainly examples of such movements of salt are often reported. Teakle and Burvill (1938) found upward movement of salt in the field in the absence of a water table. Richards, Gardner and Ogata (1956) found that in a deeply irrigated field plot of bare sandy loam soil, salt returned to the top 0-10 cm layer as evaporation proceeded over a 59 day period. The water lost by evaporation was 5.6 cm and the salt accumulating near the surface could be largely accounted for if the loss of water from below took place in the liquid phase. In addition (as a minor complication of interest) the salt distribution in the top 0-15 cm of soil was at first affected slightly by water vapour moving down from the warmer surface, condensing at a depth of about 10 cm, where the soil was cooler, and returning as liquid with salt dissolved in it. This was noted as being similar to the circulation of vapour and liquid described by Gurr, Marshall and Hutton (1952) for a closed system under the influence of a temperature gradient.

In soil subjected to seasonal wetting and drying, slow movements of liquid water (persisting for a long time) and short term (and short distance) movements under steep suction gradients probably play a role in soil genesis. Solutes can thus move towards the drying zone where water is being progressively lost by evaporation or by intake of plants if there is moist soil nearby. This probably helps the formation of carbonate, ferruginous, and other enrichments and coatings which are a feature of many soils. It is to be expected that, when plants are using the water, the distances involved would usually be small because of the rapid decrease in permeability with drying of the soil.

Although it is ordinarily true that upward movement is of little importance unless there is a shallow water table, there are thus special cases of short and long term movements of water in unsaturated soil which merit close attention in relation to soil salinity, soil genesis, water conservation in fallows, and water use by plants.

9. AVAILABILITY OF WATER FOR PLANTS

1. Permanent Wilting Point and Availability

The permanent wilting point is the water content of a soil at which plants wilt and do not recover turgidity when placed in a humid atmosphere overnight. Sunflowers are commonly used in standardized laboratory methods for determining permanent wilting point, each plant being grown in a can containing about 600 grams of soil. These plants give a clear indication of the onset of wilting and the results are reproducible. Briggs and Shantz (1912) found that this water content represented a lower limit below which any water that could be extracted by plants did not enable them to make any growth. They found this limit to be much the same for all plants in the one soil. The permanent wilting point is widely used especially in irrigation farming to represent the lower limit of water available to the crop. The upper limit of the range of available water is taken to be the field capacity whose significance and limitations have already been discussed in Section 6 (III). For farming applications, these water contents are best expressed on a volume basis—e.g., as depth of water per given thickness of each soil horizon or as depth of water within the full depth of soil occupied by the plant roots.

Permanent wilting point can be measured by the methods described by Veihmeyer and Hendrickson (1949). Also it has been found by Richards and Weaver (1943) to be closely matched by the water content at a suction of 15 atmospheres. This can be more rapidly measured and has greater significance as a soil characteristic. It is not affected very greatly by alteration of structure (Elrick and Tanner, 1955) and requires no special precautions to be taken during sampling of the soil in the field. Because the 15 atmospheres suction method has proved so satisfactory, permanent wilting point is not now measured as frequently as it used to be.

The use of permanent wilting point as a lower limit of available water for all plants has been criticised in a review of this subject by Slatyer (1957) who concludes that some plants can usefully reduce the water content of soil much below the standard wilting point as indicated by sunflowers because their osmotic characteristics are different. This is stated to be the case for plants of semi-arid and arid environments where permanent wilting point (as measured with mesophytic indicator plants) will not necessarily represent the stage at which local vegetation ceases to make effective use of the soil water. This differs from the earlier views of Briggs and Shantz (1912) on the independence of permanent wilting point from the type of plant used in measuring it. For agricultural purposes both the permanent wilting point (determined with sunflowers) and the water content at 15 atmospheres suction (determined with the pressure membrane apparatus) have undoubted value as representing the lower limit of available water for growth of crop plants. However, if permanent wilting point is not independent of the plant used in measuring it, the water content at 15 atmospheres suction is to be preferred as a soil characteristic.

Veihmeyer and Hendrickson (1927; 1955) have maintained over a number of years that water is equally available to plants for growth and transpiration between field capacity and permanent wilting point. Provided the soil is first wetted to the full rooting depth and a bit beyond, there is then little to be gained in irrigating again until it has dried fairly well towards permanent wilting point. In a number of experiments they found that yields were not adversely affected by this practice and that there was no particular advantage in more frequent watering. This concept of equally available water has influenced irrigation methods widely and has without doubt helped reduce waste of water and damage to irrigated land.

Many workers have questioned equal availability. Some of their work is perhaps open to the criticism which was early levelled by Veihmeyer and Hendrickson that the soil was not always uniformly wetted in their experiments or that the distribution of roots was inadequate. But there is evidence that increased yield of many crops will be obtained if the soil is wetted long before the permanent wilting point is reached.

There appear to be two rather strong reasons for believing that water is not equally available between field capacity and permanent wilting point.

(1) As the soil dries and the suction increases, an increasing amount of work has to be done in removing a given amount of water from it.

(2) The drier the soil the less permeable it is to water, and movement towards the root hairs will be more restricted.

The first of these reasons concerns the effect of suction on availability. The suction required to remove water at field capacity is about 0.1 atmosphere but at permanent wilting point it is about 15 atmospheres. Veihmeyer and Hendrickson (1955) claim that this reason is not very substantial because the suction anywhere within this range is small in comparison with that (of some hundreds of atmospheres) to which the vapour pressure of the air subjects the water in the leaves. Richards and Wadleigh (1952) point out however (on the basis of the work of Gradmann and Van der Honert), that the suction of the water in the leaves is not nearly as great as that indicated by the vapour pressure of the air because of the great resistance to water movement offered in the intercellular spaces, the stomata, and the non-turbulent air layer at the surface of the leaves. Veihmeyer and Hendrickson also make the point that near the permanent wilting point a small loss of water results in a great increase in suction. Thus the increase in suction from 0.1 to 15 atmospheres is minor in comparison with the rate of increase of suction on further drying. While this is true, it is not strictly relevant to what happens within the available range.

The effect of water movement on availability has received less attention than the effect of suction. This is partly because of the views that

(1) roots search out the water in the soil and

(2) movement of liquid water virtually ceases when soil is dried beyond field capacity.

Certainly as the soil dries from field capacity towards permanent wilting point the permeability of the unsaturated soil decreases very greatly and also movement in the vapour phase will be of little consequence because only small differences in vapour pressure will occur in this range. It is therefore true that the plant depends greatly upon its extending and ramifying root system for thorough exploration of the water reserves. When exploration is not thorough, moist zones have been observed in soil of high clay content not far from roots although the plant was suffering from water shortage (Work and Lewis, 1936). But it can be seen from Fig. 14 that movement of water in the liquid phase can take place throughout the available range and therefore the rate of movement over small distances cannot be neglected as a factor affecting the supply of water to the plant. In soil in which plants are actively growing, the movement of water will be over distances of up to a few millimetres (as from the centre to the outside of aggregates) and, since the change in permeability within the available range may be of the order of a thousand-fold, it is highly probable that availability will be affected by the wetness of the soil. The reduced permeability on drying will be to some extent compensated by an increased suction gradient favouring movement towards the root hairs but this compensation is accomplished at the cost of increased suction at the absorbing surface.

On physical grounds therefore it seems likely that water is not equally available from field capacity to permanent wilting point and the weight of experimental evidence with growth and yield of plants now supports this conclusion. The evidence is not all on one side because of the different behaviour of different plants, the various criteria adopted for estimating growth and yield, and the various meteorological conditions of the experiments. Also, as Wadleigh and Ayres (1945) have pointed out, any discussion of what happens over the whole available range is rather academic in some sandy soils since there is very little water left in them after a suction of about 1 atmosphere has been reached. In addition some of the anomalies must be due to the experimental difficulty of controlling water content and of measuring it accurately when it is not uniformly distributed through the soil. Comprehensive reviews of the experimental evidence have been made by Veihmeyer and Hendrickson (1950) and by Richards and Wadleigh (1952) and Stanhill (1957).

The data of Hudson and Salter (1953), some of which are given in Table 11, show that the effect of increasing suction upon reduced yield may be observed in tomatoes even at quite low suction. Taylor (1952) working more broadly over the whole of the available range, found that

the yield of field crops of lucerne, sugar beet and potatoes, decreased with increasing suction and that there was a significant relation between yield and suction.

TABLE 11

Yield of tomatoes in relation to maximum suction reached during growth of plants (data of Hudson and Salter, 1953)

Irrigation data				Average yield of ripe fruit per plant	
Suction at which water applied		Number of times water applied	Total amount of water applied	Ailsa Craig variety	Single Cross variety
cm	atm		cm	kgm	kgm
100	0.1	136	53	3.4	2.6
200	0.2	17	38	3.2	2.2
400	0.4	9	32	3.1	2.2
800	0.8	5	31	2.6	1.8

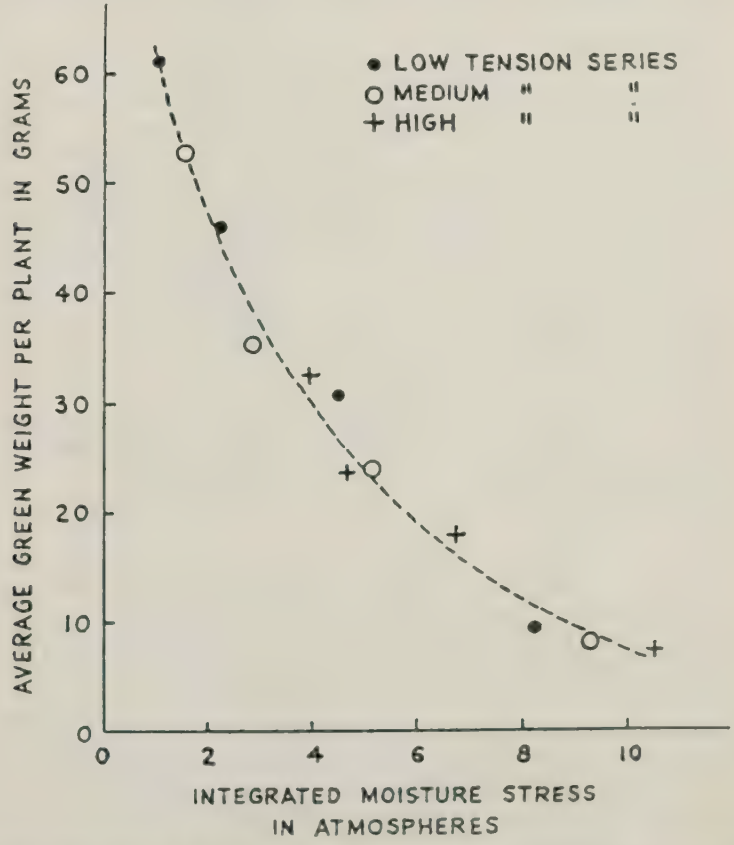

FIG. 16. *Relation of plant growth to the integrated moisture stress (or total suction) (Wadleigh and Ayres, 1945)*

To the effect of suction due to the soil matrix in opposing water uptake by plants, must be added the effect of solute suction (or osmotic pressure) due to dissolved salts if the salt content of the soil water is unusually high. Wadleigh and Ayres (1945) have shown that growth of beans decreased as the sum of these two increased. Their results are given in Fig. 16 for plants grown at three levels of matrix suction (or tension). In their diagram "integrated moisture stress" represents the average total suction (solute plus matrix suction) between irrigations. However, since permeability decreases as matrix suction increases, the effect of matrix suction on growth of plants might be expected to be greater than that of solute suction. In addition to the common effect on the free energy of the water, there is an effect on the rate of supply to the roots. Gingrich and Russell (1957) found this to be the case when they compared the effects of matrix and solute suction on root growth. Within the suction range of 1 to 3 atmospheres, growth appeared to be affected more by matrix suction than by solute suction.

A number of plant, soil and weather conditions have to be considered in deciding whether yield and quality are likely to be helped by irrigating relatively frequently or whether less frequent irrigation (letting most of the available water be used up between irrigations) will be satisfactory. Hagan (1957) has summarized these probable conditions as follows:

	Relatively frequent irrigation desirable	*Relatively infrequent irrigation possible*
Plant	Shallow, sparse, slow-growing roots.	Deep, dense, fast-growing roots.
	Fresh weight yield of vegetative organ desired.	Dry weight yields of reproductive organ desired.
	Quality dependent upon size of vegetative organ.	Harvest for content of sugar, oil, etc.
Soil	Shallow soil; poor structure impeding root growth.	Deep soil; good structure.
	Slow infiltration and internal drainage; poor aeration.	Good infiltration, internal drainage, aeration.
	Small fraction of available water held at low soil moisture stress levels.	Large fraction of available water held at low soil moisture stress levels.
	Saline soils or water.	Nonsaline.
	Fertility level high; nutrients concentrated in topsoil.	Fertility level low; nutrients distributed in profile.
	Root disease, nematodes present.	Constant water table in reach of roots.
Weather	Planted at beginning of hot dry season.	Planted well ahead of hot dry season.
	Major growth period during hot dry season.	Major growth period before hot dry season.
	High evaporation rates.	Low evaporation rates.

As an extension of this work on availability and growth, it becomes desirable also to establish the suction range which will give the most efficient use of water by plants in terms of the amount of water required to produce unit yield of a crop. Kelley (1954) has concluded from the results of field experiments that in those crops whose yields were increased by maintaining a low level of suction, there was increased efficiency of water use with regard to yield when the suction was kept low. This may be illustrated by results given in Table 12. In this experiment the "wet" treatment corresponds to a suction less than 175 cm at a depth of 12 inches, the "medium" treatment to suction less than 600 cm at 18 inches and in the "dry" treatment plants were allowed to wilt before each irrigation when the suction was approximately 800 cm at 36 inches. It can be seen that the wet treatments gave the most efficient use of water and the dry the least. Improved efficiency also resulted from increased fertility (due to phosphorus additions), a result which has also been observed by Hanks and Tanner (1952) and others from application of fertilizers to various crops.

TABLE 12

The amount of water in inches required to produce one ton of lucerne hay for various moisture and fertility treatments at Yuma, Arizona (Data of Stanberry, Converse and Haise in Kelley, 1954)

Pounds P_2O_5 per acre	Inches of water per ton of hay		
	Wet	Medium	Dry
100	10.67	11.60	12.81
200	8.79	9.11	10.49
400	7.45	8.20	8.80
600	6.82	7.45	8.86

Undoubtedly more work will be required before conclusions of wide application can be reached on the question of efficiency of water use in relation to irrigation frequency. In practice, local factors may also complicate the position, especially when there are high losses in distributing irrigation water due to channel seepage and deep penetration in permeable soil. These losses will be increased by increased frequency of irrigation.

Less work has been done on the availability of soil water for transpiration than on availability for growth of the plant. A detailed discussion by a number of contributors accompanies the paper of Veihmeyer and Hendrickson (1955) on this subject and Slatyer (1957) has also reviewed the literature. As in the case of availability for growth, there has been some controversy on whether transpiration declines with decreasing water content within the available range. This question of availability for transpiration is not of such economic importance as that on growth but it needs to be resolved because, in calculating water loss by transpiration using meteorological data (as discussed in Section 10 iv), it is necessary to know whether the rate of loss from actively growing crops is affected by the wetness of the soil within the available range. Penman (1956a) does not think the effect is ordinarily very serious. He points out that although permeability will decrease with drying of the soil, the suction gradient near the roots will at the same time increase and the rate of liquid movement to the roots may not therefore be seriously reduced within the available range. When atmospheric conditions favour a high rate of evaporation, the effect of soil water content is likely to show up (Slatyer, 1957; Bierhuizen, 1958).

Finally it may be stated that although field capacity, permanent wilting point, and available range may not be such simple soil characteristics as was once widely believed, they have served a most useful purpose in the past especially as a basis for regulating irrigation applications in many parts of the world. Recent work has produced refinements to the broad generalizations of Briggs and Shantz, Veihmeyer and Hendrickson, and others, but the generalizations themselves have been and still are of much value.

ii. Effects of Soil Composition and Structure on the Range of Available Water

The amount of water retained by a soil at any particular suction depends on the particles of which it is composed and the arrangement of those particles. As suction is increased in wet soil, the largest pores empty first followed by progressively smaller ones. Finally an ill-defined stage is reached beyond which the remaining water is held by the clay minerals rather than in the pores. The amount held in this way increases with increasing clay content and depends also on the type of clay mineral present in the soil. This effect of clay is apparent right through the available range so that the amounts of water held at both field capacity and permanent wilting point usually increase with clay content. However, the actual amount of water covered by the available range depends on structure rather than texture—the total volume of pores which are of the right size to be full at field capacity and empty at permanent wilting point.

The data of Heinonen (1954), for 129 representative soils of Finland at a depth of 5-15 cm under pasture, provide a good illustration of these relations. These soils are mainly similar in origin and contain illite as the principal clay mineral. Field capacity was determined two days after wetting and permanent wilting point was determined directly with tomato plants on some of the samples and indirectly on the rest. The soils are arranged in Table 13 into five groups of mineral soils (*a* to *e*) and a sixth group (*f*) for soils containing more than 12 per cent organic matter. The average content of organic matter in the mineral soils ranged from 4.5 per cent in group *a* to 6.5 per cent in group *e*. It can be seen from Table 13 that, in the mineral soils, permanent wilting point increases with clay content. There is a similar trend in field capacity but it is less definite because of the effect of high amounts of the silt fraction in group *c* soils and the influence of structure.

TABLE 13

Water retained in the surface layer (5-15 cm) of soils under pasture
(Average data of Heinonen, 1954, for 129 Finnish soils)

Group	Content of clay (2μ) %	Bulk density g/cm³	Permanent wilting point	Field capacity	Available water
			Per cent by volume		
a	10	1.26	11.0	25.9	14.9
b	20	1.20	15.8	35.4	19.6
c	24	1.09	17.4	39.8	22.4
d	38	1.07	19.5	37.5	18.0
e	58	1.01	25.0	41.5	16.5
f	41	0.73	21.3	44.6	23.3

The range of available water, as given by the difference between field capacity and permanent wilting point was examined in each group to find how it was influenced by particle size, bulk density and organic matter. Available water was found to increase with clay content in group *a* and with silt content in group *c*. Otherwise the most striking relations were with bulk density, especially in groups *d*, *e*, and *f*. Within each of these groups, available water increased with increasing organic matter but the relation was found statistically to be indirect. When bulk density was included in the regression equation, the influence of organic matter was largely removed and available water was found to vary inversely with bulk density in the clay soils. Heinonen attributed this to the more open granular aggregates in soils of lower bulk density.

Heinonen's data are for top soils with appreciable organic matter. Wilcox (1949) has presented a full set of data on water retention by soils in irrigated orchards of the Okanagan Valley, British Columbia. These include topsoils and subsoils all with a low content of organic matter. Wilcox found that both field capacity and permanent wilting point increased with clay content but this was not true of available water when the content of clay (<5μ) exceeded about 35 per cent. The structure probably masks any effect of clay content at that level as in the case of Heinonen's results for soils under pasture. Jamison and Kroth (1958) found that available water decreased progressively with clay content in Missouri soils. It increased with content of silt and of organic matter. Since the content of organic matter was directly related to silt content and inversely related to clay content, this apparent effect of organic matter was attributed to texture. Most of these soils had less than 2 per cent of organic matter. Regression lines are shown in Fig. 17.

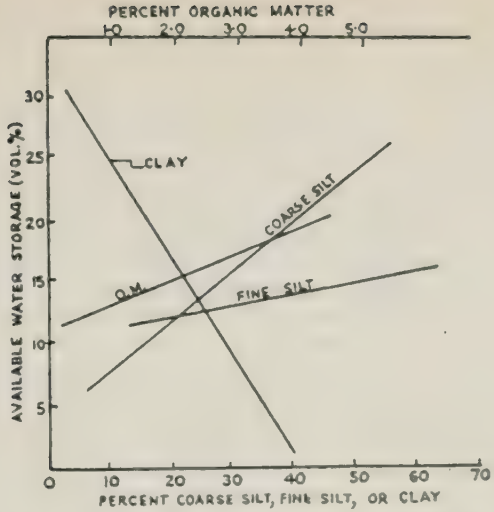

FIG. 17. *Effect of content of clay, silt and organic matter on the range of available water in some Missouri soils. Available water is here represented indirectly by the difference in water content at suctions of* 0.33 *and* 15 *atmospheres* (*Jamison and Kroth*, 1958)

Other soil constituents may also affect the range of available water by their effect on pore size distribution. El Ashkar Bodman and Peters (1956) considered that iron oxide deposited in the pores reduced the range, but Prebble and Stirk (1959) did not agree with this conclusion.

Jamison (1953) has examined the effects of various soil treatments on available water and has reviewed earlier work on the addition of organic matter and soil conditioners to soil. He concluded that available water was generally decreased by these treatments (which were otherwise favourable for increased infiltration and aeration). Consistent improvement in available water from organic matter addition was found only in coarse textured soils. These conclusions differ from those of Heinonen who found the main effects of organic matter to be in soils containing about 30 per cent or more of the clay fraction. The two points of view can perhaps be reconciled. Jamison dealt with soils to which the organic materials were added as in tillage operations. Heinonen's soils were undisturbed under pasture. In the pasture soils the organic matter is more intimately associated with the soil particles and Heinonen considers that it apparently improves the porous structure of the aggregates so that there are more pores of the right size for holding available water. He has no data on this, but his view is supported by the conclusion of Barley and Tisdall (1959) that the bulk density of aggregates was lowered when irrigated pastures were established on soil which previously had carried only sparse plant growth.

10. MEASUREMENT OF WATER CONTENT AND SUCTION *IN SITU*

The water content of soil in the field can be measured periodically by taking samples but, because of soil variability, sampling may have to be done so intensively as to be both expensive and destructive of the site. The objection will be only partly met by expressing the field water content relative to some other value which is a characteristic of the sample (e.g. moisture equivalent, water content at 15 atmospheres suction, plastic limit). To overcome these difficulties, much effort has gone into the development of *in situ* methods for measuring the changing water content and suction in soil at fixed locations.

A keen demand exists for these methods. Many field experiments with plants require periodic measurement of water content or suction. Control of water in irrigation agriculture is aided by such measurements and they are also used to follow moisture changes under pavements or buildings. The simpler and cheaper the procedure, the wider will be its use especially by the irrigation farmer; but there is also a place for the more complex procedure in the hands of the research worker. Either suction or water content measurements may be preferred depending on circumstances. A large number of methods have been tried out, but few have been accepted for routine field use. Some features of known methods are summarized in Table 14.

Kelley *et al* (1946) have reviewed the methods available up to 1946 for measuring water content or suction in the field. Literature on the subject is prolific and discussion here will be restricted chiefly to subsequent work. Reference may be made to a review by Eastwell (1953) for an extensive bibliography. Probably the most practical review of instruments suitable for routine use is that of S. J. Richards (1957).

1. Direct Measurement of Suction

The tensiometer is the only tool available for measuring suction directly in the field. A porous ceramic cup buried in the soil is connected by water-filled tubing to a manometer (Fig. 4A) or Bourdon gauge on which the pressure deficiency of the water is indicated. It is easily the best moisture measuring instrument in moist soil at suctions less than about 850 cm (pF 2.9). It breaks down at higher suctions and needs to be recharged with water when the soil again becomes wet. This nuisance of breakdown is its greatest drawback.

The range covered, although limited in terms of suction, is an important one in relation to intensive plant growth and, in sandy soils, it covers most of the water available to plants (Marshall, 1945). Its usefulness has been under-rated because of the emphasis given to the large gap in suction between its limit of operation and the suction at permanent wilting point.

Details of tensiometer design and operation have been summarized by Richards (1949) who commented on diurnal fluctuations in readings of tensiometers placed near the soil surface. Haise and Kelley (1950) have since shown these to be due to the conduction of heat between the above-ground and the under-ground parts of the tensiometer along metal tubing. Any resulting temperature gradient between the tensiometer cup and the soil causes movement of vapour and so affects the suction at the cup. In good design this effect is minimized by avoiding metal and by leading the tubing through an extended path underground.

The tensiometer could no doubt be improved for field use by the introduction of a null-indicator method to reduce the exchange of water between instrument and soil and to eliminate lag in a rapidly changing system (as has already been done under laboratory conditions). A more important step would be to develop new methods capable of covering suctions greater than 850 cm. The pressure membrane apparatus of Richards has opened this range up in the laboratory and perhaps similar methods will yet be developed for field use. For example, an indirect method might be used based on the principle that the pressure required to force air through the water-filled pores of a given porous cup depends on the suction of the soil around the cup.

II. Indirect Measurement of Suction

The indirect methods for measuring suction employ buried porous blocks which change in water content according to the changing suction of the soil water. Their use depends on the relation of suction to the weight, electrical conductivity, capacitance, thermal conductivity or other chosen property of the absorbent block. In the most widely used method, the electrical conductivity of gypsum, fibre-glass, or nylon blocks is measured.

The gypsum block developed by Bouyoucos and Mick (1940) has come into general use for field studies of both agricultural and engineering problems. It has an advantage over nylon and fibre-glass blocks in that these are very sensitive to the effect of solutes on electrical conductivity (Weaver and Jamison, 1951). The solubility of the gypsum introduces a buffering effect. Moreover, Weaver and Jamison did not find readings to be reproduced satisfactorily on successive drying cycles with nylon and fibre-glass blocks even in the absence of salt effects. On the other hand the gypsum blocks are less sensitive than nylon and fibre-glass in the wet range and are generally unsuitable for conditions wetter than pF 2.6.

Calibration of blocks (log resistance—log suction) can be undertaken in a pressure membrane apparatus as adapted for the purpose by Haise and Kelley (1946). This is best done with the blocks embedded in soil (Aitchison and Butler, 1951). Blocks made of one type of plaster in a specified way, usually give similar parallel curves. Hence if blocks from one batch are first checked according to a single point, preferably in the middle of the moisture range, it is necessary only to have a complete calibration curve on a representative sample from the group. Gypsum blocks and accessories are readily available commercially. Measuring instruments have been described by Pereira (1951), Croney, Coleman and Currer (1951) and Aitchison (1953).

The advantages of the gypsum block method are its relatively wide suction range and the ease with which it can be set up and the resistance read in the field. It suffers from a number of disadvantages. It is insensitive in wet soil and the buffering effect for salts, already referred to, does not extend beyond low salt concentrations (Richards and Campbell, 1950). Also, as noted by Bouyoucos and Mick (1947), the life of a gypsum block is limited in wet soil due to the solubility of gypsum. Thus the calibration curve may change with time and this cannot be checked upon. Further, there is considerable hysteresis between the drying (release) curve and the wetting (intake curve) for the resistance—suction relation. In this connection Tanner and Hanks (1952) found that a resistance of 23,000 ohms corresponded to any suction between 2.45 and 4.65 atmospheres (pF 3.39 and 3.67). This wide range was contributed to by variability of blocks as well as by hysteresis. Also in practice hysteresis in the resistance-suction relation of the block is not such a serious source of error since information is usually wanted under drying conditions. For this reason calibration is often done for the drying curve only.

The method has to be regarded as an approximate one and as such it is proving useful. Attempts to improve it have included the development by Youker and Dreibelbis (1951) of blocks made of fibre-glass surrounded by gypsum to extend the useful range of the gypsum block. To avoid conduction through soil as well as through gypsum, blocks have been made with concentric electrodes by Pereira (1951), Croney, Coleman and Currer (1951) and Closs and Jones (1955). This is a reasonable modification although Slater (1942) and other workers had earlier found no practical advantage in this design. Bouyoucos (1953) has improved the durability of the blocks by dipping them in a nylon resin. Buffering is said to be little affected by this treatment despite reduced loss by solution. Bouyoucos (1954, 1956) has also reduced the size of the blocks and altered the electrode shape. He claims that sensitivity in wet soil has been increased in so doing.

A method which depends on the air permeability of a ceramic cup has been tried by Kemper and Amemiya (1958). The air permeability increases with soil suction once the air entry value of the cup has been exceeded. Methods of this sort will depend upon getting ceramic or other porous cups which have a wide enough range in pore size.

Other methods for the indirect measurement of suction in the field (Table 14) have received little attention recently. Attempts made by Shaw and Baver (1939) and Haize and Kelley (1946) to develop thermal methods have shown the possibilities but have not led to their adoption for field use. One difficulty appears to be that a suitable absorbent giving the required sensitivity has

TABLE 14. Methods for Measuring Water Content or Suction *in Situ*

	Method	*Moisture Range*	*Limitations*	*Use*
1.	Direct measurement of suction. Tensiometer.	Suctions less than 850 cm (pF 2.9). Can also indicate positive pressures. (Indirect methods cannot do this.)	Needs to be re-charged if soil dries out.	Widely used especially in soil which is kept moist. Available commercially.
2.	Indirect measurement of suction using absorbent blocks.	Full range possible but actual range depends on pore size distribution of absorbent. (This applies to all methods which use absorbent blocks.)		
	i. Gravimetric.		Response may be slow because of poor contact between absorbent and soil.	Little used but further investigation of method warranted.
	ii. Electrical conductivity.			
	(a) Gypsum blocks.	Suctions greater than 400 cm (pF range from 2.6 to 4.2).	Life is limited in wet soil. Affected by moderate and high concentration of soluble salts. Approximate measurement only.	Widely used as an approximate method. Available commercially.
	(b) Nylon or fibre glass blocks.	More sensitive than gypsum blocks at suctions less than 600 cm.	Affected by soluble salts even at low concentrations.	Not used as widely as gypsum blocks.
	iii. Thermal conductivity.	Full range possible depending on absorbent.	Not fully developed as a field routine method.	Not used but warrants further study because of freedom from salt effects.
	iv. Dielectric methods.	Full range possible depending on absorbent.	Reliability not established. Electrical conductivity methods have the advantage of simpler instrumentation but are more subject to salt effects.	Not used.
	v. Air permeability of porous cup.	Full range possible depending on type of cup.	Reliability not yet established. Difficulty in obtaining cups with suitable pore size distribution; clogging of pores may prove to be a difficulty.	Not widely tried yet.

Method	*Moisture Range*	*Limitations*	*Use*
3. Water content — from measurement of related properties of soil.			
i. Electrical conductivity.	Wide range.	Highly affected by salt concentration.	Not used.
ii. Thermal conductivity.	Wide range in sand; limited sensitive range in other soils.	Not successful in soils which swell and shrink because of poor contact. Individual soils have to be calibrated.	Not widely used.
iii. Dielectric methods.	Wide range.	Reliability under field conditions not established. May require special calibration for particular cases.	Not established as a field method, but has possibilities.
iv. Neutron scattering by hydrogen atoms in soil.	Full range covered.	Not accurate near surface. Affected by high amounts of organic matter, Cl, Fe, B, crystal lattice water.	Promising method. Available commercially but expensive. Suited to measurements of progressive changes in water content at a given site.
v. Resistance of soil to penetration of a probe.	Wide range covered in clay soils.	Requires special calibration for each horizon. Restricted to soil of high clay content containing no stone.	Used to a limited extent as an approximate guide to irrigation need.
4. Water content — estimated from gains and losses of water.			
i. Transpiration calculated from meteorological records.	Applies only when water is freely available to plants.	Not applicable to estimation of water content if unknown losses occur by way of drainage or run-off.	Used as a guide to irrigation need.
ii. Measurement of vertical transfer of water vapour.	Full range covered.	Not applicable to estimation of water content if unknown losses occur by way of drainage or run-off.	Not used for routine field work. Has promise if instruments can be simplified.

yet to be developed for use with thermal methods for determining suction. Also measuring instruments have not yet been made sufficiently robust for field use. Since the thermal conductivity of an absorbent block depends on its water content and is not affected by salts, thermal methods offer some promise.

III. Measurement of Water Content

The methods for measuring suction directly or indirectly can be used to measure water content if a curve is available relating water content to suction. This may be obtained on laboratory samples provided the calibration is not affected by alteration in soil structure. Alternatively a curve can be built up by taking sufficient field samples periodically when the soil is at various suctions for measurement of water content by oven drying. In both cases the work is time-consuming because the relation has to be determined for each soil horizon represented in the field measurements.

Properties of a porous medium which vary with its water content may be measured using soil itself as the medium instead of using an absorbent block which is in suction equilibrium with soil. These properties are related to the suction of the soil when blocks are used and to water content when the measurements are made on soil. Properties of soil which have been used are listed in Table 14.

Electrical conductivity has been shown to be of little use for measuring water content, mainly because of the serious effect of electrolyte concentration. Thermal conductivity is free from this objection and de Vries (1953) has found that it works well in sandy soil. The method suffers from effects of irregular contacts in any soil subject to volume change. Also, the calibration curve relating thermal conductivity and water content differs in different soils and moreover has an unsatisfactory shape in clay soils. For these reasons thermal methods have more promise for measuring suction indirectly, using absorbent blocks.

The capacity between two electrodes placed in soil is dependent on the water content of the soil. Since the dielectric constant of water (80) is very different from that of dry soil (5) this should form a sound basis for measuring water content. Much interest was taken at one time in dielectric methods but the work done then did not reach the stage of establishing their usefulness for routine work under field conditions. Childs (1943) considered that the methods being tried at that time were likely to be unreliable because of the effects of poor electrode contact when the dielectric is leaky. It was recognized that results were quite empirical and special calibration appeared to be necessary for each soil. However, Anderson (1943) considered that the curves relating capacitance with water content in different soils were probably of a similar shape and that a single point for each soil might be sufficient for calibration purposes once the general shape was established. Use has been made of dielectric methods under laboratory conditions for measuring water content of isolated columns of sand and slate dust (Childs and Collis-George, 1950) and of surface soils *in situ* (de Plater, 1955). If a suitable method can be developed for field purposes it could also be used with an absorbent block as a suction method which should be less affected by salts than the electrical conductivity methods.

The property which hydrogen has of slowing down and scattering fast neutrons has been used by Belcher, Cuykendall and Sack (1950) to determine the amount of water in a soil on a volume basis. This principle has also been applied by Spinks, Lane and Torchinsky (1951), Gardner and Kirkham (1952), and Sharpe (1953). The use of a relatively small source of fast neutrons by Gardner and Kirkham (polonium-beryllium source) and by Sharpe (radium-beryllium) in conjunction with a boron tri-fluoride counter for the slow neutrons, brought the method within range of being economically practicable. The development of equipment by Underwood, van Bavel and Swanson (1954), Holmes (1956), and Stone, Kirkham and Read (1955) has resulted in increasing field use of the method. Holmes and Turner (1958) have published full details of a portable instrument with which they use a 5 mc radium-beryllium source lowered into a polythene-lined access hole. The calibration curve relating counting rate to water content was found to be approximately linear. This is based on data for a sand, a loam, and a clay soil and for water in bulk.

The method has the advantage of measuring the amount of water in a large volume of soil (1 cu ft or more). A restriction of the depth range of each measurement (1 foot or more at present) would be a desirable improvement for certain purposes. The method is inaccurate near the surface because of the escape of fast neutrons from the soil and results are affected by unusual amounts of organic matter (because of its hydrogen content) and of certain elements (notably Cl, Fe, and B) found in soil. Also results will not be identical with those obtained by oven-drying methods since all water molecules including those retained at 105°C will cause slowing and scattering of fast neutrons. This is not of consequence when gains and losses of water from the profile are being measured periodically at the same location.

IV. Estimation of Water Loss Due to Evaporation and Transpiration

A balance sheet of water losses (by evaporation, transpiration, run-off, and drainage) and gains (by precipitation and irrigation) will give a measure of the changes occurring in the water content of soil. In many circumstances this balance sheet could be drawn up satisfactorily if the amount of evaporation plus transpiration could be determined. Various methods have been developed for calculating this loss from meteorological records.

By considering the energy balance and the influence of wind and vapour pressure difference on the transfer of water vapour to the atmosphere, Penman (1948; 1956b) has derived an equation which enables evaporation from a water surface to be calculated from mean air temperature, mean air vapour pressure, mean wind force, and mean duration of sunshine. He has applied it to water loss from sub-irrigated turf and irrigated sugar beet in south-eastern England using an empirical factor of 0.6 to 0.8 to convert loss from a water surface to loss from a land surface. His analysis can be extended to cover transpiration, including the effects of stomatal geometry and night closure of stomata (Penman, 1952).

The evaporation, E_r, from a transpiring surface is given by

$$E_r = \frac{\Delta H_r + \gamma E_a}{\Delta + \gamma/SD} \qquad \text{- - - - - - - - - (28)}$$

where H_r=daily heat budget at surface (from incoming radiation from sun and sky, hours of sunshine, fraction reflected, and outgoing radiation).

Δ=slope of the vapour pressure-temperature curve at mean air temperature,

$E_a = f(u)\ (e_a - e_d)$,

e_a=saturation vapour pressure at mean air temperature,

e_d=actual vapour pressure of air,

u=mean wind speed,

γ=a constant,

S=stomatal factor,

D=day length factor.

The way in which S, D, and $f(u)$ may be evaluated and details of units are given by Penman (1952). Since differences in stomatal behaviour were considered to be relatively unimportant, no attempt was made to distinguish between different crops on that basis, or to allow for any effect of stomata under conditions of high evaporation. However, crop characteristics such as height, which affect the ventilation of the crop, have been considered in a particular extension of the transpiration equation to cover the case of irrigated orchard trees. By this stage the number of approximations is considerable, but the progressive steps are reasonably developed and a check against total irrigation plus rain, during four irrigation seasons at one orchard site in Australia, is shown to be satisfactory.

Other workers have developed formulae based empirically on selected climatic elements. Thornthwaite's (1948) formula for evaporation plus transpiration (which he terms "evapo-transpiration") depends on mean air temperature and latitude alone. It has been used in the control of supplemental irrigation in eastern United States where it has also been checked against soil water data (Thornthwaite, 1952). At Toronto, Sanderson (1948) found that it gave values

which were in error by as much as 40 per cent on a day-to-day basis, but which were satisfactory over a period of a week or more. The formula makes no allowance for cloud, wind or humidity and can give misleading results in other climates (Leeper, 1950; Gentilli, 1952; Mather, 1954).

Blaney and Criddle have expressed monthly evapotranspiration (or "consumptive use" of water) as proportional to the product of mean monthly temperature and monthly per cent of daylight hours of the year. The proportionality constant has different values for different crops and a correction may be introduced for humidity if records are available (Blaney, 1951). This formula has been applied in irrigated land of western U.S.A. but as pointed out by van Wijk and de Vries (1954) the irregular lag of air temperature behind solar radiation means that a method depending on monthly temperature cannot be generally applied in different regions.

The loss of water from a water surface provides the basis of other estimates of loss from vegetated land surfaces. The loss may be measured by evaporimeters or calculated from meteorological records. Tank evaporimeters are widely used but are subject to criticism on the grounds that dimensions, exposure, colour of pan, and interference from animals may seriously affect results. Prescott (1949; 1955) proposed that monthly evapotranspiration could be calculated from monthly evaporation (E_w) from the water surface of an evaporimeter using the formula $a\,E_w^b$ where b is found to have a value of about 0.75 and a probably varies from 0.6 to 2.4 according to type of vegetation (for monthly values in inches). When evaporimeter data are not available, Prescott calculates E_w indirectly from atmospheric saturation deficit (as determined from mean monthly temperature and relative humidity). This formula is used in connection with a climatic index developed by Prescott.

A number of special instruments are in use for gauging evaporation losses. These include the black pen evaporimeter of Briggs and Shantz and the black and white atmometer (Halkias, Veihmeyer and Hendrickson, 1955). The "evapotranspirometer" of Thornthwaite, described by Sanderson (1948) is an instrument which measures loss of water from wet soil through plants instead of from a water surface. Here plants are grown in a tank containing soil in which a water table is maintained (Sanderson, 1948) or, alternatively, to which water is added to the surface as required (Mather, 1954). Mather has discussed the procedure fully. A large buffer area, similarly treated as to crop and irrigation, has to be maintained around the tank.

The accuracy of any calculation of water storage in soil from meteorological data, is limited by the assumptions and approximations necessary when using such data. The soil poses some additional difficulties too, especially if run-off or drainage occurs. If a running account of the soil water content is to be kept on the basis of gains and losses, the soil has to be regarded as a simple store which is replenished by precipitation or irrigation and is drawn upon freely by transpiring plants. The capacity of the store was assumed to be 10 cm of water by Thornthwaite (1948) or it may be taken less generally, as Penman (1949) has done, to be the maximum available water which can be held by a particular soil within the depth containing or affected by plant roots. If the time at which the store is full or empty is known, the water stored (or, in practice, the water deficit) at a subsequent time can be calculated from gains and evaporative losses. Water added in excess of storage capacity is regarded as lost to run-off or deep drainage. The assumption is also made that while available water remains, the transpiration rate is unaffected by the water content of the soil or by the distribution of water within the soil. This is discussed further in Section 9.

The whole string of approximations and limitations involved in these methods is indeed so long that it might appear an unlikely way of measuring water storage in soil. But no one who has tried to measure the changing water storage in variable soils by taking samples will lightly dismiss the climatic approach. It has a definite place especially in broad scale work such as irrigation control. Work is going on actively in this field and is likely to lead to a better understanding of plant and soil mechanisms concerned with gains and losses of water. This approach is thus basically a very valuable one.

In all the above methods the water loss from the soil is calculated or measured for the case when water is assumed to be freely available to the plant. The climatic formulae can indicate large evaporative losses even when there is no water for the soil to lose. Hence the results obtained with them represent "potential evapotranspiration" in Thornthwaite's terminology. What then

of the cases where restricted water supply to the plant or the soil surface limits the rate of loss to the atmosphere? What is the "actual" rather than the "potential" loss? Much less has been done here. Staple and Lehane (1944) have dealt empirically with evaporation losses and water storage in fallowed soils in Saskatchewan. The amount lost by evaporation following each fall of rain was calculated from the expected depth of penetration and from evaporimeter data using empirical relations which had been worked out previously between evaporimeter and soil losses. A balance sheet, carried forward through the period of summer fallow, gave the amount stored at the end of the season. Empirical relations between the rate of loss of water from a crop, the amount of available water in the soil, and the rate of loss from an evaporimeter have been drawn up by West and Perkman (1953) for an irrigated citrus grove and by Butler and Prescott (1955) for wheat and pasture.

Much promise lies in methods being developed for measuring the rate at which water vapour passes upward from the soil surface. An actual measurement of water vapour transfer would meet all conditions irrespective of soil water content or type of evaporating surface. Thornthwaite and Holzman (1942), Pasquill (1950), and Rider (1957) have attempted to do this by measuring vapour pressure and wind velocity at two levels. Swinbank (1951) has developed an instrument capable of measuring the vertical transfer of water vapour over a short period of about a minute and further developments of this type of instrument are reported by Taylor and Dyer (1958). Methods of this type will find ready application if they can be made more suitable for routine use.

v. Methods Suitable for Routine Use

From the large number and the variety of methods that have been developed, it might be believed that there is a big choice open to the person who wants to select one of the methods for repeated routine measurements over one or more seasons. In fact few of them are established as routine procedures and all have their limitations. It might, therefore, be worthwhile singling out those which appear to be the most useful in the present state of their development. The selection of any particular method depends on what it is expected to do and especially on the moisture conditions in which it will have to operate. The purpose and limitations of each method are summarized in Table 14 where methods are classified according to whether they measure suction or water content. Those that are most useful and are available commercially are listed as follows in relation to the expected soil conditions and general purpose of the measurements. The method numbers refer to Table 14.

Irrigation control under continually moist conditions:
- Method 1—tensiometer.
- Method 4 I—meteorological records.

Irrigation control when the soil gets fairly dry between irrigations:
- Method 2 II—electrical conductivity of absorbent blocks.
- Plant methods—(e.g. detection of any check to the growth of fruits or shoots).

Observations of seasonal changes in wetness:
- Methods—as for irrigation control.
- Method 3 IV—neutron scattering method.

Measurement of the *quantity* of water gained or lost:
- Method 3 IV—neutron scattering method supplemented by sampling of surface soil.
- Method 4 I—meteorological records (if soil does not get too dry).

Drainage experiments:
- Method 1—tensiometer (and test wells for measuring the height of water tables).

Planning of water needs in advance of irrigation:
- Method 4 I—meteorological records (but the introduction of irrigation may later affect evaporation).

Measurements at considerable depths:
- Method 3 IV—neutron scattering method.
- Method 2 II—electrical conductivity of adsorbent blocks.

11. WATER AND SOIL PROPERTIES CONCERNED IN TILLAGE

1. Water in Relation to Consistence, Strength, and Compaction of Soil

Many of the physical properties of soil are affected by its wetness. The appearance and feel of soil depend upon how wet it is. A soil of medium or clayey texture which is hard when dry will change progressively to a friable and then to a plastic and sticky condition as it becomes wetter. These changes are so easily observed that they are often used as a guide to the wetness of soil as for example by a farmer judging the need for irrigation. Atterberg (1911) showed that remoulded soil could be dealt with as having a plastic range as defined by an upper limiting water content (the liquid limit) at which soil could just be made to flow and a lower limiting water content (rolling-out limit) at which it could just be rolled into a thread. The plastic range (or plasticity number) as given by the difference between these two limiting water contents, is found generally to increase with increasing clayiness of the soil. These Atterberg constants perform the functions of guide markers in a soil consistence scale much as field capacity and permanent wilting point do in availability. For the classification of soil in road construction they are commonly determined as a matter of routine.

A large number of methods have been developed for examining plasticity, adhesion, strength and soil deformation generally. These consistence methods have been reviewed by Bodman (1949) and practically all the properties dealt with can be shown to be dependent on suction or water content. For example the strength of a saturated block of remoulded clay varies with the water content; and adhesion of soil to metal varies with the suction. A broader discussion of the properties affecting tillage and of the contributions of Nichols to this field is given by Baver (1956). Early work with dynamometers and the work of Haines (1925) on the effect of water content on friction of a metal slider is covered by Keen (1931).

Fountaine (1954) has studied the adhesion of soil to a foreign body such as a plough share. He found that, in soil wetter than the sticky point, adhesion to a plate was given simply by the suction of the soil water multiplied by the area of contact. Some of his results are given in Table 15. When the loading in his test was done rapidly, adhesion was somewhat higher because

TABLE 15

Adhesion (g/cm^2) of Soil to a Plate (Fountaine, 1954)

Suction (cm of water)	Clay loam		Sandy loam		Sand	
	Slow	Rapid	Slow	Rapid	Slow	Rapid
3	3.1	5.2	3.1	5.7	2.3	5.9
6	6.2	14.2	5.1	9.8	4.4	8.1
9	8.6	20.9	9.1	11.0	6.4	10.2
12	11.6	20.4	11.8	12.4	11.4	12.8
15	15.0	18.3	13.5	15.7	11.9	17.7

the suction increased locally at the soil-plate joint before water could move in from the rest of the soil. If the soil is dried beyond the highest suction of Table 15 to the stage where film contact is poor, there will then be a smaller area of contact, and adhesion will fall again. Clean scouring of tillage implements is helped by low adhesion, by high cohesion (which increases with suction), and by good aggregation. Fountaine and Payne (1954) have considered these and the implement factors affecting scouring.

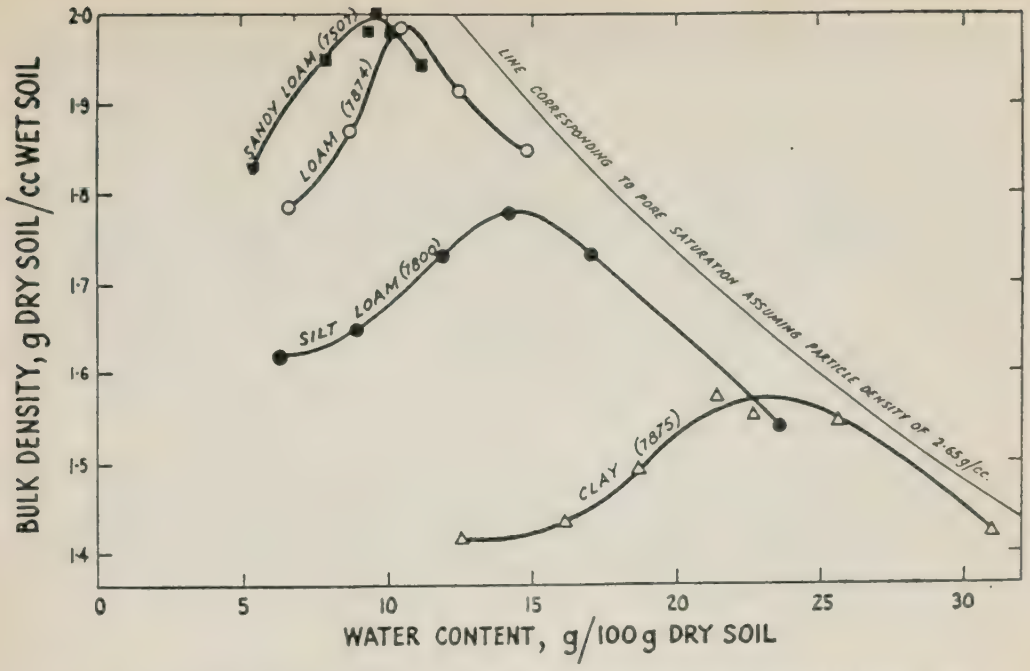

FIG. 18. *Compaction of soil in relation to its water content (Compaction done by method of Proctor,* 1933)

The importance of compaction in farm operations has been stressed by Nichols, Cooper and Reeves (1955). It can cause reduced aeration, increased run-off, and plough soles which resist the penetration of roots as well as water and air. It results from the pressure of tillage tools and from the traffic of machines and animals. Söhne (1953; 1958) has presented data showing how porosity is reduced when various loads are applied to a loam soil at a range of water contents. The lower the water content, the greater is the pressure required to reduce the porosity to a given value. For example, to obtain a porosity of 0.42 the pressure required on his air-dry, intermediate, and wet soil respectively is in the proportions 50 : 9 : 1.

The relation between water content and compaction, when the soil is pounded by a falling hammer in a standardized fashion, is illustrated in Fig. 18. It can be seen that, for each soil, an increasing water content at first assists the rearrangement and closer packing of particles. The water acts as a lubricant between particles and aggregates and there is also a reduction of the shear strength of the aggregates themselves so that flattening increases at areas of contact between aggregates in the manner observed microscopically by Day and Holmgren (1952). Above a certain optimum water content for maximum compaction, additional water prevents close packing by occupying the remaining pore space. (Because of this, the optimum water content is found to decrease and the maximum bulk density for a soil increases if an increased amount of work is done in compaction.) For a given method of compaction, the highest density is reached in soil that has a wide distribution of particles from coarse to fine (as in the sandy loam of Fig. 18). In this case the fine particles fill up the gaps between the coarse. High density is not reached in soil made up of coarse or fine particles only. The method on which Fig. 18 is based (Proctor, 1933) is that commonly used by highway engineers and consists of ramming soil at various water contents in a standard fashion into a standard cylinder. The method is not ideal for tillage studies since pressure is applied in a different manner. However, Weaver and Jamison (1951) found that compaction under tractor is related to water content by the type of curve

provided by the Proctor method. They concluded from their trials that compaction may be considerable near the water content which is otherwise desirable for ploughing. In clay soils, the effect of mechanical compression upon density is usually small in comparison with the effect of the maximum desiccation and shrinkage which the soil may have undergone in its history. This is discussed in Section 3 (III).

The deformation of soil by machines and by animals involves shearing of aggregates as well as a general compaction of the soil and, since in agriculture the breaking up of aggregates in this way occurs principally under moist and wet conditions, it is referred to by Bodman and Rubin (1948) as a puddling process. Bodman and Rubin have examined the puddling of beds of aggregates packed in an annular ring to which both normal and tangential stresses could be applied. They have expressed the puddlability of the soil as the change in volume of the bed per unit of work done in causing the change. Puddlability was found to increase with increasing water content whether puddling is by normal stress alone or by tangential and normal stress combined.

Under very wet conditions, the treading of stock or the passage of vehicles can cause deformation by plastic flow without any increase occurring in the bulk density. The aggregated soil is then reduced to a puddled state and permeability and other properties are much affected although there is no loss of porosity. Gradwell (1956) describes conditions such as these in pastures during winter in New Zealand where the air space porosity (which was low any way) was decreased little by severe treading of stock.

II. Stability of Aggregates Exposed by Tillage

Water can have marked effects upon soil exposed by tillage operations to the action of rain. Hénin (1938) reviewed and examined the probable effects of water upon the stability of exposed aggregates. Some of the points he made can be summarized as follows:—

1. If water enters rapidly, the stresses that are set up by swelling of clay (due to unequal wetting of different parts of an aggregate) can weaken or break it.

2. On the hypothesis of Yoder (1936), water entering rapidly into dry aggregates compresses air in the smaller pores and this tends to break open the aggregate. Slow wetting allows the small pores to fill by capillarity first and air can later escape easily from the large pores with less damage.

3. The drier the soil the greater is the breakdown on wetting rapidly.

4. The presence of organic matter slows up wetting and thus increases stability.

5. Swollen aggregates offer less resistance to the mechanical action of rain.

6. Strength is given to the aggregate during drying by orientation of clay particles by their linkage through exchangeable cations and oriented water molecules (as proposed by Russell, 1934), and by bonding with oxides and organic materials.

Many of these points have been examined experimentally during the last 20 years and within this time there has come the spectacular development of synthetic polymers capable of increasing the stability of aggregates in water. In general, the points made by Hénin are supported by later work. There is now rather more emphasis on swelling than on trapped air as a cause of breakdown if the soil contains sufficient clay particles. Low (1954) found that breakdown still occurred in some soils even when they were evacuated before wetting and Brewer and Blackmore

(1956) and Panabokke and Quirk (1957) were able to attribute breakdown principally to swelling in most soils of sufficient clay content.

The marked effects of initial dryness and rapid wetting on breakdown have been measured by a number of workers. They may be illustrated by the data in Table 16 from the work of Russell and Tamhane (1940) for an unmanured, arable soil of fairly low stability. These two causes of breakdown are considered by Cernuda Smith and Vicente-Chandler (1954) to be just as important as falling or running water. They suggest that attention should be paid to the effect of covers to protect the surface against complete sun drying and rapid wetting.

TABLE 16

Effect of initial water content and rate of wetting on percentage of aggregates larger than 1 mm that are stable in water (Russell and Tamhane, 1940)

Method of wetting	Initial water content, % by weight	
	17 (field moist)	1.7 (air dry)
Slow wetting by capillarity	54	5
Fast wetting by immersion	36	2

It appears then that when soil becomes exposed by tillage operations, the aggregates are weakened and destroyed by the rapid wetting of dry aggregates by raindrops. Also the kinetic energy, amounting to 4,500 ergs for a drop of 3 mm diameter at terminal velocity, has to be dissipated by aggregates which are already in a weakened condition when moist. A number of workers have examined the effect of rain on the water relations of soil at the surface and the notable work of Duley and Kelley (1939) has already been mentioned in this connection. McIntyre (1958) examined the permeability of a silty loam soil after subjecting it to 3 cm of rain applied artificially during 30 minutes with drops whose energy was equivalent to that of 3 mm raindrops. At the immediate surface a thin skin of compact soil about 0.1 mm thick was formed with a hydraulic conductivity, k, of only 5×10^{-7} cm/sec. Below this there was a zone a few millimeters thick into which fine particles had been washed. Here k was found to be 5×10^{-6}. In contrast to these low values for the two zones constituting the crust, k was equal to 10^{-3} cm/sec in the unaffected soil below.

There is ample evidence of progressive deterioration in the water stability of aggregates in soil which is regularly tilled and this decline appears to be associated with a decline in the content of organic matter. The organic matter is believed to protect the aggregate by reducing the rate of wetting (Emerson, 1954) and by adding to the wet strength by its bonding properties (Emerson, 1951; Panabokke and Quirk, 1956). There are, however, good reasons for tilling soil, some of which are touched on incidentally in earlier sections and accounts of which are given in detail by Russell (1950), Middleton (1952), Baver (1956) and Greacen (1958). Although tillage may have the adverse effects on aggregates mentioned here, it improves structure for particular purposes as for example in the preparation of a seed bed. Vilensky (1945) in a comprehensive paper on structure and tillage in relation to soil moisture went so far as to suggest that improved aggregation could result from tilling the soil in a special way at the right water content so as to form small

pieces of it into dense stable aggregates. The most favourable stage for tillage was between the plastic limit and the sticky point (which is the water content at which kneaded soil will just adhere to external objects). The higher the pressure that could be exerted on the aggregates, the lower was the optimum water content required for high density and stability—a condition which is true of any compaction process. Vilensky was criticised by Kachinsky (1949) for claiming that desirable aggregates could be formed by manipulating and compacting soil and certainly there is widespread evidence of progressive deterioration of aggregates in intensively cultivated soils. When aggregates are broken down they are not easily reformed and the result is poorer permeability and aeration and, in regard to the tillage process itself, a narrower range of moisture conditions within which tillage can be undertaken. A well aggregated soil behaves under tillage as though it were more sandy than it actually is and this allows a wider range of water content over which the soil can be worked without being either too hard or too sticky.

LIST OF PRINCIPAL SYMBOLS

Some of the following symbols have been used with a meaning other than the main one given below but if so, they are defined at the point of use in the text. This is the case also with symbols not included in the list.

Å	Ångström unit (10^{-8} cm)
c	water content, volume of water per unit volume of material
c_w	water content, mass of water per unit mass of dry material
D	coefficient of diffusion of a gas in a porous material
D_o	coefficient of diffusion of a gas
E	evaporation
g	acceleration due to gravity
h	length of water column representing the suction
H	hydraulic head
i	amount of water entering soil in time t
k	hydraulic conductivity
k'	intrinsic permeability
l	length
log	logarithm to base 10
ln	logarithm to base e
N	normal concentration
p	vapour pressure of water in soil
p_o	vapour pressure of water in bulk
P	hydrostatic pressure
q	quantity of gas diffusing across unit cross-sectional area in unit time
r	radius
R	gas constant
S	surface area of particles per unit volume of porous material
t	time
T	absolute temperature
v	volume of water flowing across unit area in unit time
w_m	weight of gas adsorbed for completion of first molecular layer
Δ	difference
δ	small difference
ε	porosity (fraction of total volume of porous material occupied by pores)
ε_g	gasfilled porosity
η	viscosity
μ	micron (10^{-4} cm)
ρ	density of water
ρ_a	bulk (or apparent) density of porous material
σ	surface tension of water
φ	gravitational potential
Φ	potential ($\psi + \varphi$)
ψ	capillary potential

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